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DIASTROPHIC BEHAVIOR AROUND THE BIGHORN BASIN[1]

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ABSTRACT

A striking feature of the mountain belt around the Bighorn Basin is the reversal of the direction of overthrusting and the asymmetry of folding in transverse segments of the ranges. Not only do these segments of alternating southwest and northeast thrusting characterize the ranges, but the asymmetry of each continues in minor anticlines well into or completely across the adjoining basins. Five such transverse strips are described.

Deformation was that of the Laramide foreland in which the pre-Cambrian rocks and structures, relatively thinly covered, were a more important determining factor than in the geosynclinal folding farther west. The greatest crustal shortening was northeast-southwest. In the basement complex there were two types of yielding: (1) horizontal shearing along two sets of near-vertical fractures, each at about 45° to the shortening, breaking the foreland in places into blocklike units; and (2) vertical movements giving ranges and basins with various individual departures from the general Rocky Mountain alignment. In the adjustment to space requirements under horizontal compression, transverse strips of the ranges and basins moved differently.

The minor folds have a regional northwest-southeast trend more or less regardless of the trends of the main ranges and basins which were outlined earlier. This minor folding was relatively shallow; near the surface easiest relief was upward, and the folds were aligned normal to the greatest shortening. Although the basement deformation did not greatly influence the trend of these lesser folds, it did determine the direction of their asymmetry.

INTRODUCTION

Many of the areal geologic studies by different members of the Yellowstone-Bighorn Research Association, under the general direc-

[1] Presented in part before the Geological Society of America, December 27, 1939.

tion of Dr. W. T. Thom, Jr., have been published during the last few years; and there have appeared, in addition, individual contributions on various special problems of Middle Rocky Mountain geology. At intervals several progress reports have brought out some of the more generalized results of this co-operative research. As considerable work still remains to be done before a comprehensive summary report can be undertaken, another report of progress on some of the larger aspects of the work in hand now seems in order.

REGIONAL RELATIONS

The isopach map of the Rocky Mountain region from Montana southward shows the greatest thicknesses of Paleozoic and Mesozoic formations to occur in a north-south strip extending through westernmost Wyoming and eastern Idaho south into Utah (Fig. 1). These sediments amount to more than 25,000 feet in this strip which obviously represents a long-lived geosyncline. This geosyncline apparently developed more irregularly than the Appalachian trough, forming several especially pronounced basins, one of which, centered near the corner of Idaho, Utah, and Wyoming, is indicated by an accumulation of over 40,000 feet of sediments.

Eastward, the Paleozoic and Mesozoic sediments thin rapidly, though irregularly, to about 10,000 feet in the Beartooth-Bighorn region. There is, however, no well-defined eastern border of the geosynclinal belt; if one wishes to draw a dividing-line between the trough and the platform region to the east, it must be located arbitrarily. Within the platform region, as in the geosyncline, one notes a prevalence of sinking basins which received more sedimentation than neighboring areas. This irregular behavior presumably reflects the influence of earlier structural developments.

Late in the Jurassic the Nevadian orogeny involved large portions of the Pacific states and near-by parts of Idaho and Nevada. East of the strongly deformed belt was a strip of the Cordilleran geosyncline running through eastern Idaho and westernmost Wyoming, as already noted, which remained essentially undeformed. During the Cretaceous, detritus from the western mountains continued to be added to the thick accumulation of Paleozoic and Mesozoic sediments already there. In later Cretaceous time an approach toward

geosynclinal conditions reached eastward to the Beartooth-Bighorn region, which previously had been an interior platform area. Com-

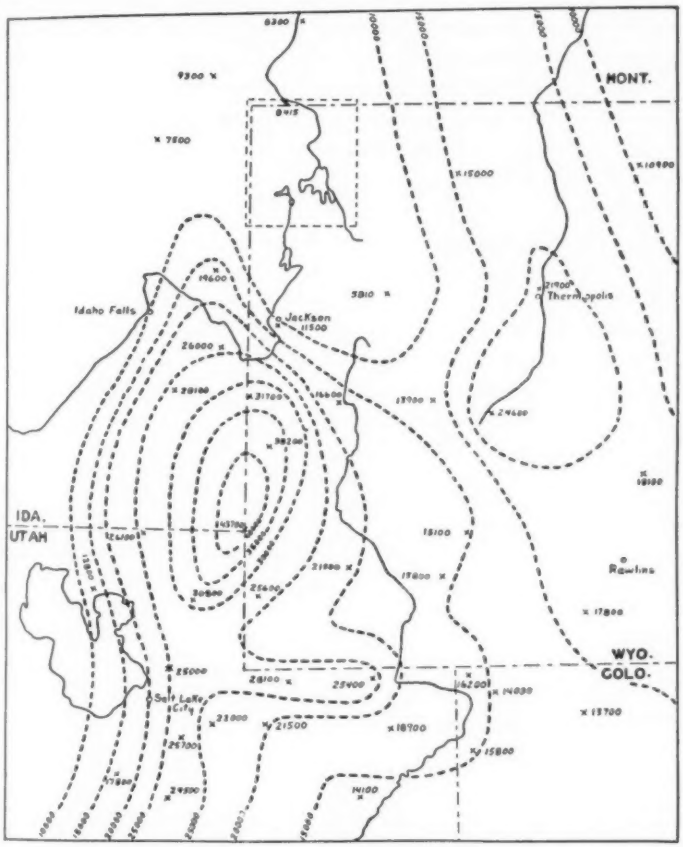

FIG. 1.—Isopach map showing thickness of Paleozoic and Mesozoic strata in western Wyoming and adjacent areas. From Leland Horberg.

pared with the earlier rock systems, the Cretaceous strata in the Bighorn region are relatively thick, bringing the total Paleozoic-Mesozoic column up to 8,000–15,000 feet. But this thickness is

much less than that of the corresponding column farther west in the geosyncline proper.

Crustal activity progressed eastward, and the Laramide compression folded the thick, weak strata of the geosyncline proper into anticlines and synclines of the Appalachian type, though the individual ranges were less continuous than in the Appalachian system. The folding was accompanied by overthrusting of major proportions. The results were especially pronounced where the sediments were exceptionally thick.

In easternmost Idaho, just west of the Wyoming line and extending into Utah, is the remarkable Bannock overthrust described by Richards and Mansfield.[2] Its sinuous trace has been followed for about 270 miles, and the horizontal displacement is known to be more than 12 miles and may be as great as 35 miles. This overthrust was located near the deepest part of the trough, where the sediments were thickest. Sediments on the westerly rise from the bottom of the trough were thrust eastward approximately as far as the axial line. Not far east of this, on the easterly rise from the bottom of the geosyncline, are the Crawford, Absaroka, Darby, Jackson,[3] and several lesser thrusts (Fig. 2). While the present traces of these do not follow exactly the curvature of the isopach lines, they nevertheless show considerable relationship to them, particularly in curving around the northeast side of the pronounced "deep" mentioned above. This local "deep" has obviously exerted some influence upon the faulting and its location. But the influence of the conditions and forces responsible for the general Rocky Mountain trend is also apparent, leading to the conclusion that the curving trends of the individual faults in this strip of overthrusting have resulted from a combination of these two controlling factors.

East of the geosyncline, whose thick strata were strongly folded and thrust faulted with much crustal shortening, is what may be called the "foreland area" of the Laramide orogeny. Its pre-Lara-

[2] R. W. Richards and G. R. Mansfield, "The Bannock Overthrust: A Major Fault in Southeastern Idaho and Northeastern Utah," *Jour. Geol.*, Vol. XX (1912), pp. 681–707.

[3] Leland Horberg, "The Structural Geology and Physiography of the Teton Pass Area, Wyoming," *Augustana Library Pub. No. 16* (1938). Pp. 86.

mide strata were thinner than those to the west, and it deformed differently. Over extensive portions of it the pre-Cambrian crystal-

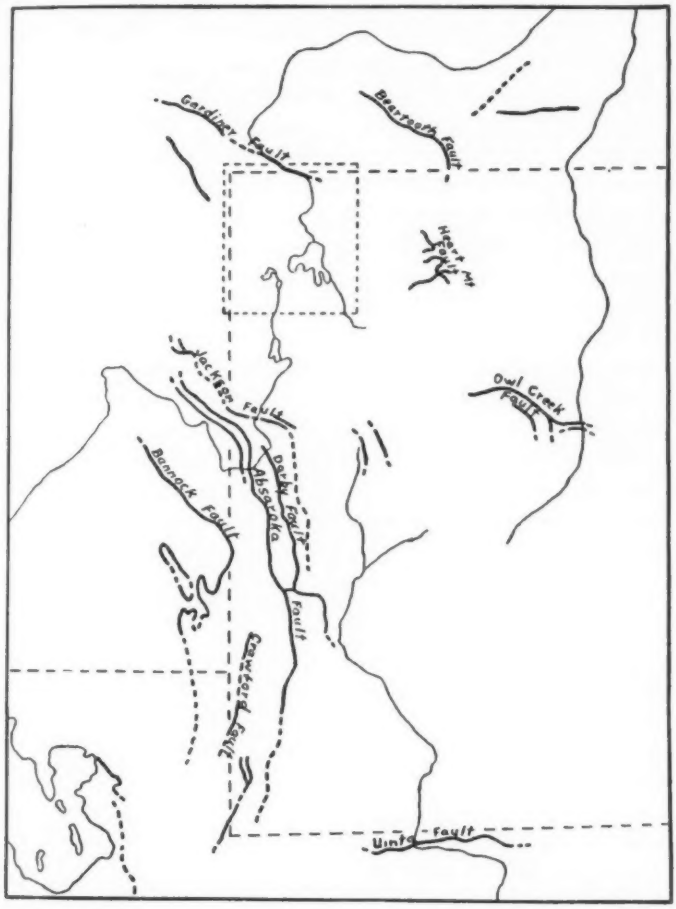

FIG. 2.—Prominent thrust faults in western Wyoming and adjacent areas. From Leland Horberg.

line complex now appears at the surface, whereas the formations now exposed in the geosynclinal folds and fault slices of western Wyo-

ming and adjacent Idaho are chiefly of Paleozoic and Mesozoic age.[4] When the Laramide orogeny began, the basement crystalline rocks of the Beartooth-Bighorn region reached in general to within less than 15,000 feet of the surface and beneath large areas to within only 10,000 feet of it. The sedimentary cover was too thin a sheet to dominate and determine the detailed behavior of the foreland area in the ensuing deformation, as the much thicker series of strata did in the geosyncline, where the basement rocks were deeply buried. Instead, the pre-Cambrian crystallines exerted the decisive control in the Bighorn–Beartooth–Wind River platform or foreland area. Instead of long anticlines and synclines of Appalachian type, there was predominantly an uplifting of scattered block-shaped or oblong ranges and a sinking of basins. Crustal shortening was only moderate.

Areally the basins are greater than the uplifts, but the two opposing features developed together and must be treated together as parts of the same orogenic manifestation. Our study is primarily of the Bighorn Basin, one of the most typical of these intermontane basins, and its encircling ranges, among which are some of the most strongly developed mountain uplifts of the foreland area.

THE BIGHORN-BEARTOOTH DEVELOPMENT

The Bighorn Basin is nearly surrounded by a horseshoe-shaped belt of uplifts which, though developed in separate units, are related parts of the whole. These units, beginning on the northwest, are the Beartooth, Absaroka, Owl Creek–Bridger, Bighorn, and Pryor mountains and the much feebler Nye-Bowler lineament. Some of the larger of these mountain units consist of subunits, or segments, each of which deformed with a certain individuality of its own. Characteristic of this individuality is a one-direction asymmetry of structural features which persists throughout each individual segment but (in both the Beartooth and Bighorn ranges) is in the direction opposite from that of the adjoining segment on either side. The reversal of direction of structural asymmetry is one of the outstanding peculiarities of the tectonic development.

[4] Farther north, in Montana and the Canadian Rockies, the enormously thick Belt series of sediments reaches the surface in the geosynclinal folds, as do also Proterozoic strata in the continuation of the strong folding and faulting to the south in Utah.

REVERSAL OF STRUCTURAL ASYMMETRY

Let us take a swing around the horseshoe belt of ranges for a general picture of the striking reversal in the direction of overturning of folds and overthrusting on faults which alternates between southwest and northeast in a succession of segments.

The Beartooth Range, in the large, appears to be a single unified uplift; but in detail it is composite, consisting of subunits—the North Snowy block and the South Snowy block composing the northwestern half of the main range and the Beartooth Plateau block the southeastern half, plus the Canyon Mountain appendage which extends the North Snowy structures a short distance northwestward and the Rattlesnake Mountain appendage which continues the range to the southeast. By their different behavior in deformation these subunits constitute three transverse segments of the range. (1) The North and South Snowy blocks and Canyon Mountain appendage (northwestern portion of the range) were uplifted most strongly on the southwest, while their thrust faults dip northeast. They may be taken together as a northwestern transverse segment of the Beartooth Range characterized by southwestward overturning. (2) The Beartooth Plateau block (southeastern portion of the main range) was uplifted most on its eastern and northeastern sides and exhibits thrusting toward the east and northeast. The direction of overturning and overthrusting in this southeastern transverse segment of the Beartooth Range is thus the opposite of that in the northwestern segment of the range. (3) The Rattlesnake Mountain appendage, extending southeast from the Beartooths, is characterized by southwest overturning, which again is the reverse of that of the adjoining Beartooth Plateau block.

The Laramide structures of the Absaroka Range, next southeast in the horseshoe curve, are largely concealed beneath Tertiary volcanic rocks, but the asymmetrical anticlines in the Bighorn Basin alongside it have their steeper limbs on the southwest. The same is presumably true of the Absaroka Range, which would thus continue the southwest turning manifest in the Rattlesnake appendage. This conclusion is strengthened by the line of isolated Paleozoic and pre-Cambrian areas just within the western margin of the volcanic cover to the south of Yellowstone Lake. They show strong uplifting on the

southwest side of a range which was either the Laramide Absarokas or else a continuation of the Wind River Mountains. The Owl Creek–Bridger Range at the south end of the curving horseshoe is an unsymmetrical uplift having gentler north slope and steeper south slope with overthrust faulting outward from the Bighorn Basin. Although with some curvature of trend, the asymmetry is harmonious with that of the Absarokas and the Rattlesnake appendage of the Beartooths.

The Bighorn Range on the east side of the basin consists of at least three segments of alternating symmetry.[5] (1) The southern segment, which curves to meet the Bridger Range, was pushed up most on the southwest and has the steep limbs of its minor folds in the same direction, thus continuing the strip of southwest deformation around the south end of the Bighorn Basin. (2) The central Bighorns have been most strongly uplifted on the east and thrust east-northeast toward the Powder River Basin. (3) In the northern Bighorns the conditions are again reversed. The outcrops of pre-Cambrian rocks (structurally the highest portion of the broad segment) are near the southwestern flank of the range, and the bordering thrust faults, dipping under the range, are along that margin. (4) In the northernmost tip of the Bighorns the uplift lessens rapidly, and this portion of the range is essentially a small structural pleateau bounded by monoclines on each side, that on the northeast being slightly the greater. Considered in conjunction with the more marked asymmetry of the near-by folds in the Powder River Basin, it would seem that the northernmost Bighorns properly belong to a transverse segment in which the dominant surficial movement has been directed northeastward. This is a reversal of the direction shown by the northern Bighorns. Segments 1 and 3 of the Bighorns thus have their structures inclined in one direction; segment 2 and, less decisively, segment 4 in the opposite direction.

All this forms an astonishing pattern of alternating directions of structural asymmetry. Yet this is only a part of the picture. Not only do these transverse segments of one-direction asymmetry extend completely across each mountain range, but the symmetry of

[5] W. H. Bucher, W. T. Thom, Jr., and R. T. Chamberlin, "Geologic Problems of the Beartooth-Bighorn Region," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934), pp. 168–76.

each segment continues unchanged into, and in some cases across, the adjacent basins. The region is thus divided into transverse strips of one-direction asymmetry (Fig. 3).

FIG. 3.—Tectonic sketch map of the Bighorn–Beartooth–Wind River area showing the strips of reversed asymmetry. Arrows indicate direction of overturning of folds.

STRIPS OF ONE-DIRECTION ASYMMETRY

For convenience of reference the transverse strips of one-direction asymmetry of this region will be designated by numbers, beginning at the northwest corner of the area.

STRIP 1

Strip 1 comprises the two Snowy blocks of the western Beartooths and the Canyon Mountain appendage, together with the near-by

portion of the Crazy Mountain Basin on the northeast. As already noted, the two Snowy blocks have been uplifted most on the southwest, and the South Snowy block has been overthrust[6] southwest on the Gardiner fault.[7] On the northeast flank of the North Snowy block the thrusting has likewise been southwest up the gentle slope of the block or outward from the Crazy Mountain Basin. Within this strip the relative surficial movement has been southwestward.

STRIP 2

The southeastern half of the Beartooth Range (Beartooth Plateau block) has been thrust strongly toward the east and northeast. Near its east-facing front there are no minor folds in the Bighorn Basin, but in the eastern half of the basin the observed anticlines have steep northeast limbs and gentler southwest limbs. Continuing eastward, across the structural trend, one finds the four major blocks of the Pryor Mountains tipped up highest at their northeast corners. Perhaps with the Pryors, Strip 2 should properly end, as it does with the pronounced Sheep Mountain anticline some distance to the south southeast; nevertheless, although the northernmost Bighorns are not far from symmetrical, the steeper northeast limbs of near-by folds in the Powder River Basin just east of the northern tip of the Bighorns suggest a further eastward continuation of Strip 2 in that latitude. This transverse strip of steeper northeast sides thus crosses the Beartooth Range, the Bighorn Basin, and the Pryor Mountains, with some evidence for its continuation farther east. Throughout this strip the relative surficial movement was toward the northeast.

STRIP 3

Adjoining on the southeast is Strip 3 in which the asymmetry is in the opposite direction. The Rattlesnake Mountain appendage of the Beartooths is tilted up most strongly on the southwest and in places has been thrust southwest along bordering thrust faults. Likewise, the minor folds in the western portion of the Bighorn

[6] In this description of the strips of asymmetry the possibility of underthrusting and underturning is held in abeyance. To make the discussion as simple and the picture as clear as possible, the relations are all described as overthrusts and overturns.

[7] C. W. Wilson, Jr., "Geology of the Thrust Fault near Gardiner, Montana," *Jour. Geol.*, Vol. XLII (1934), pp. 649–63.

Basin facing the Rattlesnake appendage the Absarokas and the Owl Creek–Bridger Range show turning toward the southwest away from the Bighorn Basin. The Owl Creek–Bridger Range, south of the Bighorn Basin, is, for the most part, very unsymmetrical with gentle dips on the north slope and steep ones on the south side. Near the Wind River Canyon through the Owl Creek Mountains a low-angle overthrust fault not far from the southern border has brought the Cambrian onto steeply inclined Triassic beds.[8] The thrusting was directed south outward from the Bighorn Basin. The central section of the Owl Creek Mountains is more nearly symmetrical. The granite area is bounded on both sides by faults, but even here the apparent dominant movement was toward the southwest. The southern Bighorns were pushed up most on the southwest, and the same was true of the near-by anticlines in the Powder River Basin, as well as those around the southern end of the Bighorn Basin. Asymmetry in the same direction continues the strip southwest across the Wind River Basin and characterizes also the Wind River[9] and Gros Ventre[10] ranges beyond. This belt of southwestward overturning thus completely crosses three main ranges and two large basins, as well as a portion of the Powder River Basin. It is the longest and largest of the strips.

STRIP 4

The asymmetry is again reversed in the central Bighorns, the highest portion of the range. The Paleozoic strata on the western side exhibit moderate dips; on the eastern flank they stand more nearly vertical and, in places, are cut by reverse faults with northeastward thrusting. The folds in the neighboring portion of the Bighorn Basin are similarly turned toward the northeast. In the Powder River Basin in front of the eastward thrusts of the Bighorns there was, however, little development of minor folds, as if the mountain thrust faults had here sufficed to take up the crustal shortening. The thrusting in Strip 4 has been toward the northeast.

[8] J. R. Fanshawe, "Structural Geology of Wind River Canyon Area, Wyoming," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. XXIII (1939), pp. 1439–92.

[9] C. L. Baker, "Geology of the Northwestern Wind River Mountains, Wyoming," unpublished manuscript.

[10] Vincent Nelson and Victor Church, manuscripts in preparation on the structural geology of the western and central portions of the Gros Ventre Range.

STRIP 5

The northern Bighorns reverse the direction of the central portion of the range, indicating surficial thrusting toward the southwest, with turning of the anticlines near by in both the Powder River and Bighorn basins in the same direction. This strip, however, does not extend far into the Bighorn Basin. Beyond, farther out in the basin, the folds are turned the other way in a weakly deformed part of Strip 2 south of the Pryors, which possibly involves also some of the effects of basin flexing (see p. 714).

These strips of unidirectional deformation show another dynamically significant feature. In the area *in front* of a strongly overthrust mountain front, as east of the central Bighorns or east of the southeastern Beartooths, minor folds are conspicuously absent. Apparently the upward and outward escape of the rocks along such a faulted mountain border relieves the stress in front of it, exerting actually a protective effect on the near-by portion of the adjoining basin.

BOUNDARIES BETWEEN STRIPS OF OPPOSITE ASYMMETRY

The boundary lines between the strips of oppositely directed asymmetry in Figure 3 were drawn primarily from the asymmetry of the anticlines and uplifts shown on the geologic map, with less regard for mapped cross structures. Are there distinct structures which coincide with these boundaries? There must have been adjustment of some sort between the belts. One might expect such adjustment to have been accomplished in several possible ways: (1) by differential horizontal movement concentrated along a single fault or a zone of faulting; (2) by some bending of the rocks in a shift zone; (3) by a gradual accommodation throughout a moderately wide intermediate zone; or (4) by some combination of these processes.

ON THE RANGES

Between the central and southern segments of the Bighorn Mountains is the Tensleep fault, the most prominent transverse fault in the range. It is a boundary fault between Strips 3 and 4. The upthrow is on the north, the central Bighorns being the most strongly raised portion of the range. In places the throw of the fault is as

much as 1,200 feet, according to C. W. Wilson, Jr., who has made a special study of it.[11] The surface of slip is nearly vertical, but dipping for the most part toward the south. Drag along the fault shows it to be normal, but reversal of drag along one short section of it indicates later reversal of slip movement and suggests later compression. Wilson thinks that undercrowding of the southern unit against the northern side has caused a turning of the fault surface so as to give here a north dip of 85°. An overcrowding of the high-standing central Bighorn segment against the lower southern Bighorn segment would also produce the observed turning.

Wilson has reason to think that the Tensleep fault was formed early in the Laramide orogeny and that the thrust fault on the west side of "the Horn" (south of the Tensleep's eastern extremity) was related to it in time but continued to function longer. As brought out in other publications, the early phase of the Laramide orogeny was dominantly differential uplifting of the ranges, whereas horizontal thrusting became more important in the later phase. Although the Tensleep fault continued straight east across the range in the early phase of orogeny, separating the central from the southern Bighorns, the thrusting in the later phase at the southeast corner of the central Bighorn block was southwestward. A triangular area at the southeast corner of the central Bighorn segment has minor anticlines on the east flank of the range with steeper sides toward the southwest and some faults dipping east northeast. This triangle was detached from the central block during the later phase of the orogeny by the Crazy Woman cross fault, recognized by Demorest,[12] and behaved like the adjoining southern segment. Although first a part of one segment and later moving with the other segment, this small triangular area, between the Tensleep and Crazy Woman faults, is here considered as a part of the central Bighorns and placed in Strip 4 because the asymmetry of the larger mountain subunits developed in the earlier stages of the orogeny is taken to be more important than the asymmetry of the minor folds developed subsequently.

[11] C. W. Wilson, Jr., "The Tensleep Fault, Johnson and Washakie Counties, Wyoming," *Jour. Geol.*, Vol. XLVI (1938), pp. 868–81.

[12] Max Demorest, "The Structure of the East Central Part of the Bighorn Mountains," unpublished Doctor's thesis, 1938, Princeton.

From Darton's geologic map transverse faulting is to be suspected between the central and northern Bighorn segments along the boundary between Strips 4 and 5. Demorest has recognized what he has termed the "Tongue River fault" in the postulated position of the boundary between the segments on the eastern side of the range. In the middle of the range, however, the boundary bends to the southwest and continues across Shell Creek to the Bighorn Basin with gradual, rather than abrupt, change in the general structure on the west side of the range.

Along Wolf Creek, south of Dayton, near the northeast corner of the central segment of the Bighorns, Darton has noted some minor folding (with faulting) in the flanking Paleozoic strata which is turned toward the southwest[13] and is thus out of harmony with the tilting of the segment. Some other structures likewise indicate a surficial creep up the steeper northeastern side of this mountain segment toward its northern corner. Two significant facts are to be noted: (1) The flanking Paleozoic beds here dip northeastward less steeply than farther southeast, where the turning is consistently northeast throughout. (2) This portion of the mountain flank adjoins the northern Bighorn segment, whose thrusting is toward the southwest.

Here is a triangular area, bounded on the north by the Tongue River fault and on the south by the strong tear fault separating it from the most prominent eastward salient of the central Bighorns, which bears strong resemblance to the triangle already noted at the southeast corner of the central Bighorn segment. We may note that the faults on the south sides of the two triangles look like representatives of the east-west lines of shearing which characterize our whole region (see p. 702), whereas the faults making the northwest sides of the triangles appear to be tension fractures normal to the long dimension of the range. The triangular blocks outlined by these faults behaved differently from the main segment on each side and did not develop in their entirety the characteristic structures of either one, but instead show an overlapping of the two directions of asymmetry. The two indicate some difference in pattern between the earlier and the later movements.

[13] N. H. Darton, "Geology of the Bighorn Mountains," *U.S. Geol. Surv., Prof. Paper 51* (1906), pp. 93–95; also Demorest, *op. cit.*

Likewise, the composite nature of the northernmost Bighorns seems to suggest opposite direction of yielding in the two episodes of deformation—toward the southwest with the northern Bighorn segment in the earlier one and northeast with Strip 2 in the later one. The primary uplifting here was much weaker than farther south, as the range flattens out northward, and no clear-cut boundary line developed.

The boundary between Strip 2 and Strip 3 passes along Pat O'Hara Mountain, which trends east and west between the southeast extremity of the Beartooth Range, and the Rattlesnake Mountain appendage. Pat O'Hara Mountain is bordered on its south flank by a strong high-angle fault running east and west. The whole structure is one of adjustment between blocks of uplifting, and combined with this accommodation there has been also horizontal adjustment between strips of opposite asymmetry.

IN THE BASINS

Faults along the boundaries between strips of different structural asymmetry have not been recognized to any great extent in the basins, though sufficient search has not yet been made for them. Perhaps the accommodation was more dispersed in the weak sedimentaries seen in the basins than in the stronger rocks now exposed by extensive denudation of the ranges.

The Nye-Bowler structural lineament, from the Stillwater River on the north side of the Beartooths eastward to the line of division between the northern and southern Pryors, does, however, give evidence of differential horizontal shifting.[14] The oblique echelon faults, strung out along the lineament, are thought to have resulted from tearing of the surficial sedimentary beds by horizontal movement on a long deep-seated fault, or shift zone, which located the lineament. The northeast-southwest trends of the individual faults obliquely across the lineament say that the area south of the lineament moved east relative to that to the north of it. Likewise the southern Pryors have been more strongly tipped up at their northeast corners and more strongly thrust northeast than the northern Pryors, which are on the north side of the prolongation of the Nye-Bowler linea-

[14] C. W. Wilson, Jr., "Geology of the Nye-Bowler Lineament, Stillwater and Carbon Counties, Montana," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. XX (1936), pp. 1161–88.

ment.[15] This is all in harmony with the structural asymmetry of the whole eastern Beartooth–Pryor–northernmost Bighorn strip for which has been postulated northeastward surficial movement. The Nye-Bowler lineament, however, seems to indicate also differential movement within Strip 2. Somewhat stronger horizontal shifting in the same direction is revealed by the remarkable Lake Basin belt of echelon faults much farther north. This will be discussed later.

A shorter zone of oblique echelon faulting extends nearly due south from the southeast corner of the Pryor Mountains toward Lesser Sheep Mountain, one of the larger anticlines in the Bighorn Basin.[16] The individual faults trend northeast-southwest. This, interpreted in the same way as the other belts of oblique tension faulting, would indicate that the area immediately west of the fault zone had shifted north relative to that on the east side of the zone. This zone of faulting coincides with a part of the boundary between Strips 2 and 5. On the east side of this north-south zone the surficial thrusting has been toward the southwest, whereas on the west side the shove from the Bighorn Basin has been toward the northeast—in each case oblique to the trend of the zone. The northward shifting of this portion of the Bighorn Basin immediately south of the Pryors was an important adjustment involved in tipping up the northeast corners of the four Pryor blocks. This shift, combined with the eastward shift of the south side of the Nye-Bowler lineament, gives a reason for the fact that the two southern Pryor blocks were tipped up more strongly at their northeast corners and more strongly faulted on their eastern sides than the two northern Pryor blocks which rise immediately to the north of the Nye-Bowler lineament.

TYPES OF BLOCK BEHAVIOR

Tilting is very characteristic of the large uplifted mountain subunits, while asymmetrical anticlines prevail on their flanks and in the basins. The direction of tilting of the mountain blocks as well as the inclination of the much smaller asymmetrical folds on their borders and in the basins, is harmonious throughout each transverse

[15] D. L. Blackstone, Jr., "Structure of the Pryor Mountains, Montana," *Jour. Geol.*, Vol. XLVIII (1940), pp. 590–618.

[16] C. J. Hares, unpublished studies.

segmental strip. Both the larger structures and the smaller structures of each strip apparently must have developed under a control which had some factors in common throughout the whole strip. The behavior along the borders of uplifts and basins will next be considered.

A. CHARACTERISTIC RELATION WHERE MOUNTAIN SUBUNIT SLOPES GENTLY TOWARD ADJOINING BASIN

On the gently sloping side or "backslope" of an uplifted mountain block the Paleozoic and Mesozoic sedimentary beds have formed minor folds turned toward the crest of the range. Whatever thrust faults have developed from the splitting of such folds dip away from the range. It is clear that the sedimentary cover beds have been forced some distance up the slope of the range. In riding up the slope the upper beds have moved more than the lower ones, producing inclined secondary anticlines and in places additional translation along thrust faults. Such behavior is particularly well shown on the north slope of the North Snowy block of the Beartooth Range, southwest of Livingston, where the structures have been mapped in detail by Lammers.[17] Here the splitting of anticlines in a succession of slice faults was far advanced before the deformation ended. Not only were the sedimentary beds urged up the slope of underlying crystalline rocks of the North Snowy block, but the pre-Cambrian rocks were also involved in the slice faulting. The conclusion seems justified that the slope between basin and upwarp was a more decisive factor in governing the structures than the differences between crystalline and sedimentary rocks.[18] Whether the underlying pre-Cambrian is everywhere involved in the minor structures or not, this characteristic backslope behavior of the sedimentary cover is general throughout the whole region. The chief variation is in how far the processes have progressed.

B. CHARACTERISTIC RELATIONS WHERE STEEP SIDE OF MOUNTAIN SUBUNIT FACES A BASIN

The steep sides of the mountain blocks vary from (*a*) simple flexures with high-angle faults (as in some of the Pryor blocks) to

[17] E. C. H. Lammers, "The Structural Geology of the Livingston Peak Area, Montana," *Jour Geol.*, Vol. XLV (1937), pp. 268–95.

[18] *Ibid.*, p. 293.

(*b*) overturned anticlinal limbs and overthrust faults in various stages of development (as in two of the Beartooth blocks, in the central Bighorns, and in the Owl Creek Range). Rattlesnake Mountain,[19] the Five Springs section of the Bighorns,[20] and the southeast Pryor block[21] represent intermediate stages. In (*a*) horizontal yielding has been least relative to the uplifting. In (*b*) the horizontal compression has been more effective, and the early upbowing has given sufficient relief and steepness of slope between the uplifted block and the adjoining basin for the basin margin, under further compression, to be forced under the edge of the mountain block and the latter to bulge out above as an overturned anticline. With still further compression the overturned limb of this fold has split along a portion of the mountain front, and the upper part has moved outward toward the basin on one or more thrust faults. This is one of the most characteristic type structures of the region. The mountain block shows squeezing-in below and flaring-out above. The bordering thrust faults vary from ordinary reverse faults with moderate slip where the bulging above was not great to low-angle overthrusts with notable horizontal displacement where a greater rotational strain developed. Great difference in elevation and a steep slope between the basin and the range have been important factors in causing the low-angle faulting.

Low-angle thrust faulting has been best developed where two conditions are met: (1) The mountain block is one of the most strongly uplifted ones of the region, as shown by its exceptionally large area of pre-Cambrian rocks exposed at the surface, i.e., the Beartooth blocks and Central Bighorn block. (2) The margin of the block has a pronounced bend with convex curvature facing the basin (northeast front of the Beartooth Plateau block and the northeast side of the Central Bighorn block). In each case the highest part of the range is near the faulted margin. The basin immediately in front is deep near the range, and one may observe also that relatively few minor anticlines are to be seen in this part of the basin.

[19] G. Duncan Johnson, "Geology of the Mountain Uplift Transected by the Shoshone Canyon, Wyoming," *Jour. Geol.*, Vol. XLII (1934), pp. 809–38.

[20] C. W. Wilson, Jr., "A Study of the Jointing in the Five Springs Creek Area, East of Kane, Wyoming," *Jour. Geol.*, Vol. XLII (1934), pp. 498–522.

[21] Blackstone, *op. cit.*

In the process of crustal shortening the basin margin, where the front of the range is very abrupt and the rise great, has been thrust under the range border. The rotational strain developed under these conditions has given fracturing at a relatively low angle. The portion of the range margin where the bend is of sharpest curvature (generally the middle portion of the long side of the individual block) has experienced the greatest displacement on the low-angle fault. This is natural enough, since the flaring of the upper part of the mountain block was outward in the convex bend and the prominence of the salient itself has been increased by the outward flaring of the upper part of the block. Where the bend is particularly pronounced, the most conspicuous transverse tear faults of the region are to be seen. Here (eastern corner of Beartooth Plateau block,[22] for example) the overriding portion of the mountain mass, as it spread out in an arc of increasing circumference, developed tension concentric with its arcuate front and radial, vertical tear faults naturally developed. As displacement on the low-angle fault continued, there was differential horizontal movement along the vertical tear faults, some sections advancing farther than others, thus giving the broken east fronts of the Beartooth Plateau and Central Bighorn blocks (Figs. 3 and 4).

In the case of the Gardiner thrust the uplift of the South Snowy block behind it was great, but the curvature of the margin was only moderate. There was less displacement on the fault than along the Beartooth Plateau front, and the angle of faulting was not so low. For the Owl Creek overthrust the uplift of the range behind was less than in the other examples, and the curvature of the margin, though notable, was on a relatively small scale. The overthrust broke at a low angle, but it did not form the border structure of the range in quite such typical fashion as in the other cases. The basin in front was deep.

C. ADJUSTMENT AROUND THE CORNER OF A BLOCK

The deformational behavior around the corner of a block is best displayed on the northwest corner of the North Snowy block of the Beartooth Range (Lammers, 1937). The gentler slope of this tilted

[22] E. L. Perry, "Flaws and Tear Faults," *Amer. Jour. Sci.*, Vol. XXIX (1935), pp. 112–24.

block faces northeast. With surficial movement from the basin on the north the sedimentary beds have been crowded up this slope, forming asymmetrical folds and thrust faults along which southwestward overriding has taken place. The block itself, lifted high at an early stage, has acted as a superficial buttress confronting this movement. Around the northwest corner of the buttress and in the

FIG. 4.—Beartooth east front near Red Lodge, Montana. On bordering thrust fault pre-Cambrian of the mountain has overridden slightly overturned Paleozoic sequence from Cambrian to Mississippian. Another lower thrust fault, which has placed Mesozoic beds on the early Tertiary Fort Union, is concealed by slope wash. The transverse (Willow Creek) tear fault in the distance has given the Paleozoic palisades an offset of about ½ mile. Photo taken from near a second (Mount Maurice) tear fault which has caused an offset of about 2 miles.

much lower terrain farther west, the southwest shove of the sedimentary rocks was progressively less impeded by resistance from the buttress. In consequence, the folds and faults curved around the corner and advanced farther south in the lower country than upon the slopes of the buttress, forming pronounced arcs. Where the southward advance was greatest, low-angle overthrusts and imbricate structures were strongly developed.[23] Accommodation in this

[23] D. C. Skeels, "Structural Geology of the Trail Creek–Canyon Mountain Area, Montana," *Jour. Geol.*, Vol. XLVII (1939), pp. 816–40.

differential forward movement, which gave the arcuate trends, was accomplished by shift zones[24] and tear faults, the west side of each shift zone and tear fault being offset to the south relative to the east side. The buttressing effect of the North Snowy block is properly attributed by Lammers more to its superior elevation than to the greater resistance of its pre-Cambrian crystalline rocks, though the fact that these relatively stronger rocks rose higher in the buttress than in the surrounding area was doubtless also a factor.

NARROW ZONES OF FOLDING AND ECHELON FAULTING

A peculiar feature of the northern part of the Bighorn Basin and of the country to the north of it is the prevalence of narrow zones of east-west monoclinal or anticlinal folding and echelon faulting. From south to north these are the Nye-Bowler lineament, the Lake Basin zone, and the Cat Creek anticline. The most striking of these is the Lake Basin belt, which lies well out in front of the main chain of the Rocky Mountains. In a 56-mile stretch of this belt northwest of Billings, Montana, are ninety more or less parallel faults trending northeast-southwest obliquely across the belt.[25] The belt continues east of the Yellowstone River and extends a few miles beyond the Bighorn River. The individual faults are at very steep angles and are obviously the result of tension. Monoclinal folding parallels the belt and not the individual faults. The Big Coulee–Hailstone dome, bordering the belt of faulting on the north close to its western end, is elongate northwest-southeast in the Rocky Mountain trend, or normal to the strike of the individual faults. The fault and fold pattern is identical with that obtained experimentally in a sheet of clay by moving one portion of the clay sheet horizontally with respect to the rest.[26] The fracturing may also be compared to the diagonal crevassing along the side of a valley glacier. Such crevasses extend in from the margin obliquely upstream toward the middle of the glacier and are known to be due to faster ice movement in the middle

[24] Lammers, *op. cit.*, pp. 281–83.

[25] R. T. Chamberlin, "A Peculiar Belt of Oblique Faulting," *Jour. Geol.*, Vol. XXVII (1919), pp. 602–13.

[26] A. E. Fath, "The Origin of the Faults, Anticlines, and Buried 'Granite Ridge' of the Northern Part of the Mid-Continent Oil and Gas Fields." *U.S. Geol. Surv. Prof. Paper 128-C* (1920), pp. 75–84.

of the glacier than along the sides. Tension from the differential movement splits the ice.

It seems certain that the Lake Basin faults trending northeast-southwest obliquely across the long east-west axis of the belt were caused by the area immediately south of belt of faulting moving eastward relative to that lying to the north. It is natural to think that this differential horizontal movement has occurred along the strike of a steep fault, or zone of shearing, in the underlying basement complex. Relative eastward movement of the south side would first cause some stretching of the weak Cretaceous and earliest Tertiary strata near the surface, followed by tension tearing in a strip of short oblique faults over the line of deep-seated shearing. This deep-seated line of weakness dates from long before the Laramide deformation, for the fault belt, as noted by Thom, is closely coincident with a belt of local stratigraphic change. Already in Paleozoic times this was the locus of differential crustal movement.

The Nye-Bowler structural lineament, parallel to the Lake Basin fault zone and much closer to the mountains, has much in common with it. A deep-seated east-west fault beneath the lineament is likewise postulated.[27] Outbreak of lavas and ejectamenta along the western portion of this line of weakness near the Beartooth front in Judith River times is taken to indicate the beginning, or at least a very early stage, of the Laramide orogeny. Also relatively early, when strong vertical movements caused uplifting of the range and sinking of the Bighorn Basin, a monoclinal fold developed over the basement fault. Near the Beartooth Range at the west end of the lineament and in the vicinity of the Pryor Mountains at its east end, the high side of the fold was on the south in conformity with the near-by mountain uplifts. But in the much longer middle portion of the lineament across the Bighorn Basin the high side of the fold was on the north, because the basin which was sinking lay mostly to the south. The difference in elevation here was sufficient to cause the accumulation of thick sediments on the south side and thin deposits on the north side. Later, horizontal compression increased and lateral movement along the buried fault produced the

[27] C. W. Wilson, Jr., "Geology of the Nye-Bowler Lineament, Stillwater and Carbon Counties, Montana," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. XX (1936), pp. 1161–88.

tearing in oblique, echelon faults. The south side of the lineament moved east. The faults are more regularly en echelon in the eastern half of the belt than in the western, where igneous activity was most pronounced, though their development is far less striking than in the Lake Basin belt. In later Fort Union times (Tongue River stage), when the horizontal compression was particularly strong, the older monoclinal fold was arched and given its local anticlines and domes. According to Wilson, the lateral movement along the basement fault continued the echelon faulting.

THE MINOR FOLDS

In addition to the ranges and basins, which are the foreland structural features of the first order, there are many anticlines and synclines of a second order of magnitude which may be conveniently termed the "minor folds." We have already made their acquaintance to some extent in considering the strips of asymmetry and the types of block behavior. They are most conspicuous on the flanks of the ranges and the sides of the basins, owing, in part, at least, to the fact that the ranges have been stripped to the granite over large areas while thick Tertiary fill covers the middle portions of the basins.

To the present writer, their outstanding characteristic is that they have nearly the same trend over the whole region. Although there is local variation between about N. 30° W. and N. 60° W., their prevailing trend is around N. 45° W. The main ranges and basins, or larger units, on which the minor folds are superimposed are considerably more irregular in trend. Some geologists have called the arrangement of these lesser folds en echelon. But to the writer it would seem, rather, that in this Rocky Mountain foreland area the minor folds appear en echelon simply where their general regional trend is inclined to a range or basin somewhat differently oriented.

INFERRED NATURE OF THE MAJOR PROCESSES

GEOSYNCLINAL BEHAVIOR

The geosynclinal theory of mountain-building requires the presence of a geosyncline filled with essentially concordant sediments before the process of folding becomes operative. If one supposes the

sinking of the trough to be due largely to lateral compression, there arises the difficulty of seeing how much downward movement can be accomplished without producing tectonic deformation and greater unconformities than those which occur. Some geologists (as, for example, Bucher[28]) go so far as to assume that the formation of the sediment-receiving geosynclinal depression is the result of tensional stresses in the crust, which are thought to alternate more or less periodically with compression. Most geologists believe, however, that the processes of geosynclinal sinking and orogenic compression are different phases of one and the same continuous process. They look to contractions due to rearrangement of material deep within the earth in favor of greater density; or to changes in the state or behavior of matter in consequence of the enormous pressures at great depths, as may be suggested from an extrapolation of some of Bridgman's experiments; or to processes based on thermal changes; or to other, as yet unknown, factors. Under this view, one may suppose that the geosyncline was localized by greater mobility in that belt which permitted greater downward movement there than elsewhere.

Whatever may be the cause of the long, relatively undisturbed sinking of the geosynclinal belt, eventually the chain of orogenic movements begins. As was indicated, above, there can be little doubt that the superficial deformation is an accompaniment of a strong downbuckling of the deeper crust. Two possibilities present themselves:

1. The geosynclinal belt may be the active agent of deformation, sinking rapidly because of a greatly accelerated downdraft produced by some such causes as are mentioned above, drawing downward with it the adjoining parts of the crust, and thus causing the marginal tracts to crowd against the trough of thick sediments, under the general conditions of shrinkage as a basic postulate.

2. The geosynclinal belt may be the passive agent, being forced down by the pressure exerted on the pre-existing trough by the closing-in of an actively shrinking crust. In both alternatives, general shrinkage is an essential condition, since an active downdraft

[28] W. H. Bucher, *The Deformation of the Earth's Crust* (Princeton University Press, 1933), pp. 139–48.

along the geosynclinal belt in an inactive crust would inevitably cause extensive tension along the borders, of which there is no sign in the typical region concerned.

Whatever may have been the ultimate cause, the crust on both sides of the long geosyncline was crowded strongly inward against it in the natural adjustment by crustal shortening. Because of variations of materials, structures, heat, and stresses within the earth, the incrowding may have been notably greater from one side of the geosyncline than the other. Movement from the same side, however, need not have dominated throughout the entire length of the geosyncline. Reversal of conditions along its length is possible. If notable movement from both sides was inward and downward, the resulting folded belt would show overthrusting outward on both flanks and would extend deepest in the middle portion, on the wedge principle.[29] Although two-sided, the mountain system might nevertheless be quite asymmetrical in cross section. If one side crowded more strongly than the other, the expected result would be a folded belt whose structures were mostly inclined in one direction.

FORELAND BEHAVIOR

In Utah, in the southeast corner of Idaho, and in southwesternmost Wyoming the long folds and fault traces in the geosynclinal belt trend approximately north and south. Northward, just before the visible Laramide structures pass beneath the later volcanic cover in the Snake River downwarp, the trend lines curve around to the northwest. This bend, which is located just southwest of the Teton Range, appears to have the nature of a salient, the concealed structures west of Yellowstone Park presumably curving back again into more nearly north-south alignment to continue into the geosynclinal deformation of northwestern Montana. It looks as if the incrowding from the western side, south of the 43° parallel, was directed nearly due east or essentially normal to the geosynclinal belt. North of that latitude it was directed east-northeast. But in the foreland to the east in Colorado, Wyoming, and southernmost Montana, the prevailing trend of the structures is noticeably different. Although

[29] R. T. Chamberlin, "Isostasy from the Geological Point of View," *Jour. Geol.*, Vol. XXXIX (1931), p. 20, Fig. 3.

having considerable irregularity in the details of basins and ranges, the general regional structural trend there is northwest-southeast. In general, the difference in the two trends is about 30°. In the critical area southeast of Jackson, Wyoming, where the Gros Ventre uplift of the foreland abuts the geosynclinal folds of the Hoback Range, the trends of the latter, however, curve to the northwest, as already noted (p. 676), into harmony with the Gros Ventre structures. The Gros Ventre area crowded against the geosynclinal belt from the northeast. Thus the structures suggest more nearly direct crowding from the west and more oblique crowding from the northeast. The curving trend of the geosyncline itself has been a factor in this.

Going a step farther in the consideration of oblique surficial movement, there are two possiblities. The first is that of direct northeast-southwest compression in the foreland. The second appeals to differential horizontal shifting of crustal masses, such as is seen in California along the San Andreas rift and elsewhere, in which the Pacific side was moved northwest, relative to the eastern side. Short folds, more or less en echelon, have developed near the rifts in consequence. Prominent surface rifts like those in California have, however, not been observed in this portion of the Rocky Mountain foreland. So also, the foreland folding was more regional than that localized near rifts in California. Still, in Mead's familiar experiment of subjecting rubber coated with plastic wax to a rotational strain, folds were produced in the wax by pushing two opposite sides of his frame in opposite directions, and yet the horizontal shifting was not concentrated along distinct rifts.[30] In that experiment the resulting folds were rather short, overlapping one another. Similarly, if the western portion of our area were shifted northward sufficiently, relative to the eastern portion, or if the eastern portion were shifted strongly southward relative to the western portion, the differential movement might perhaps have developed the succession of short, scattered mountain uplifts, which display in some places an echelon arrangement of the minor folds on their flanks. Van Pelt suggested such an explanation a decade ago for the minor echelon

[30] W. J. Mead, "Notes on the Mechanics of Geologic Structures," *Jour. Geol.*, Vol. XXVIII (1920), pp. 505–23.

folds in the foothills of the Colorado Front Range which run obliquely to the axis of that north-south range.[31]

Figure 5 illustrates how such a process would operate. Given a square wire screen *ABCD* (like that described by Leith[32]) which is 10 inches on a side. Hold the left-hand side *AB* immovably and

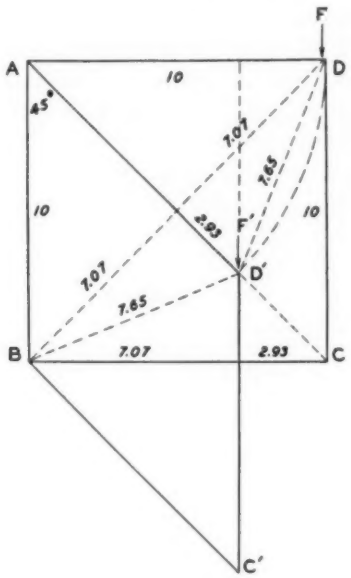

FIG. 5.—Deformation of square screen under strong rotational strain

push the right-hand side *DC* down until the right angle *DAB* is reduced to 45° (*D'AB*). The square *ABCD* has become the rhombus *ABC'D'*. By this change in shape the area of the screen has been reduced 29.3 per cent. The diagonal *BD* of the original square was 14.14 inches in length; this has become the diagonal *BD'* of the rhombus whose length is only 7.65 inches. The screen has been shortened 6.49 inches along this line. This shortening of 6.49 inches

[31] J. R. Van Pelt, unpublished seminar report, 1927. In July, 1939, the same idea occurred independently to W. H. Bucher, who was much impressed with its possibilities.

[32] C. K. Leith, *Structural Geology* (1923), Figs. 4, 5, and 6.

has resulted from shifting the corner D to D', a straight-line distance of 7.65 inches. The amount of shortening is thus *less* than the distance which D has been shifted in producing it. A similar result is obtained if sides AB and DC are both moved (in opposite directions) so as to give the same total relative shift. If folds develop from such reduction in plan area, their trends at first tend to be at right angles to BD; but with progressing distortion they are rotated to become normal to BD', as is well known.

The folding, of course, takes the orientation indicated because, although the force was actually applied parallel to DC, the resultant of the compression, or the *effective* compression, which causes the reduction in area operates at first in the direction DB and gradually shifts to $D'B$. In the wire-screen device of Leith the effective compressive stress which produces the distortion arises from the applied force modified by the cohesion of the screen. Because of the nature of the screen the individual wires parallel to its edges neither lengthen nor shorten appreciably, but instead are forced to come closer together. In the thick rubber sheet of Mead, the rubber is stretched in one direction and pulled in at right angles to this. Changes are produced in the third dimension also. In the earth the cohesion of the rocks should operate somewhat similarly.

In the type of deformation just discussed, shearing has taken place along both sets of planes of maximum shear. For comparison and contrast let us consider what may be called "card-deck shear," in which slip occurs along only one set of shearing-planes. The cards in a stack may be slid sidewise so that the square $ABCD$ becomes the rhomboid $A'BCA$ (upper diagram, Fig. 6). But unlike the distortion of the screen, this change in shape involves no reduction in area. Each individual card remains unchanged, having been merely shifted in position relative to the other cards.

Considering the working hypothesis that the foreland folding was produced by the horizontal couple postulated above, we might suppose some card-deck shear to have taken place, though this would have been subordinate to the rubber-sheet type of yielding. To the latter would be attributed reduction in ground-plan area and the resulting folding. But even on this hypothesis the folds would take their northwest-southeast alignment because the *effective* compres-

sion, though arising from a north-south force couple, operated along the northeast-southwest lines. The couple is one possible way of developing the northeast-southwest compression which was responsible for the folds. But, assuming a north-south force couple, one would expect that, if the region exhibited much strike slip along steep faults, the slip likewise would be dominantly in a general north-south direction. Such is not the case in this area. On the other hand, there is evidence of much east-west shifting along steep

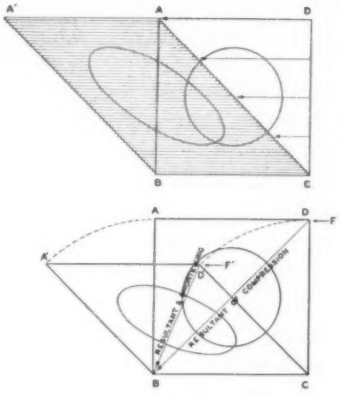

FIG. 6.—Upper diagram illustrates card-deck shear; there is no reduction in area of cross section. Lower diagram, for comparison, represents screen deformation; in this case the area has been reduced 29.3 per cent.

faults or shear zones. The significance of this will be discussed later. The other possibility is direct, regional, northeast-southwest compression across the foreland area.

Our next problem resolves itself into two distinct subproblems: (1) what outlined the less regular, larger units (ranges and basins); and (2) what fashioned the more regular-trending minor folds which show much independence of the ranges and basins.

DEVELOPMENT OF RANGES AND BASINS

During the earlier stages of the Laramide orogeny, the most conspicuous deformation of the Beartooth-Bighorn region was sinking basins and rising near-by ranges. The basement crystalline rocks,

mantled by only 10,000–12,000 feet of sedimentary strata, were the most important ones in determining the nature of the yielding to the stresses. Comprising various intrusions and occurring in complicated structures, they presumably varied much more from place to place, and far more irregularly, than the sedimentary sequence along the length of the geosyncline. It is, therefore, natural to expect a more irregular structural pattern to develop in the Bighorn–Beartooth–Wind River area than in the geosyncline. The crystallines were also strong rocks, such as would naturally fracture in block pattern.

Oblong and oval are perhaps the most characteristic shapes of the individual uplifts. In many places their borders for considerable distances are nearly straight lines in two prevailing directions—one nearly east-west, the other nearly north-south. These lines, together with a strong tendency for the ranges to follow the general Rocky Mountain northwest-southeast trend, but with some notable departures, are the outstanding characteristics of the larger areal pattern. Ranges in the general Rocky trend normal to the regional shortening are the natural expectation; hence, the various departures from this and the occurrence in many places of north-south and east-west border lines, oblique to both the Rocky trend and the shortening, must be of prime significance. What outlined such ranges with basins between many of them? Two leading causes, probably working in conjunction, naturally suggest themselves: (1) pre-Laramide structural control, such as faults and the attitudes of other structures in the basement together with notable differences in competency of the basement rocks (batholiths versus schists, etc.); and (2) vertical Laramide fracturing in conjugate north-south and east-west sets of shear planes due to lateral shearing in the northeast-southwest crustal-shortening process.

STRUCTURES AND COMPETENCY VARIATIONS IN BASEMENT

The more this region is studied, the more evident it becomes that conditions in the underlying complex of pre-Cambrian rocks have played an important part in governing the Laramide deformation. As early as 1919, horizontal differential movement along a deep-seated fault, or zone of slippage, was recognized as the most likely explanation of the Lake Basin belt of echelon faulting in the

sedimentary cover. Somewhat later, other lineaments and features farther north in central Montana were found to indicate important control by the crystalline basement.[33] More recently the Nye-Bowler lineament has been interpreted along the same general lines as the Lake Basin belt of faulting.

In the summer of 1933 Ernst Cloos and Hans Cloos applied the methods of granite tectonics in a prelinimary study of the Beartooths, Bighorns, and Black Hills to determine whether there be any observable relation between the pre-Cambrian structures of those uplifts and the later Laramide deformation.[34] Their general conclusion was that, "although differing individually in every detail, the three structural units seem to indicate coincidence between pre-Cambrian doming and their present outlines." Selecting for investigation the plunging southeast end of the Beartooths, they found the strike of the platy textural elements within the pre-Cambrian to have a north-south to northwest-southeast trend parallel to the longest axis of the present range. The linear elements plunge toward the south exclusively and follow the direction of the later uplift, but with a much steeper dip within the crystallines than in the younger sedimentary cover. The Laramide Beartooth uplift has thus followed the pre-Cambrian structure lines around the southern end of the range and accentuated the older structure in the basement.

The Drs. Cloos also concluded that the present longest axis of the Bighorn uplift coincides with the long axis of a pre-Cambrian upwarp:

> The Bighorn uplift contains a granite core, which constitutes the present highest parts, partly surrounded by a mantle of schist. . . . . Bordering faults are parallel to the primary granite structure, and flexures often transgress it. Basic dikes indicate an updoming of the pre-Cambrian, with its culmination within the granite core.
>
> The Black Hills present a complete pre-Cambrian dome with a granitic center and a schist mantle. The long axis coincides with the present axis of the uplift. Schists dip away from the granite.

[33] W. T. Thom, Jr., "The Relation of Deep-seated Faults to the Surface Structures of Central Montana," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. VII (1923), pp. 1–13.

[34] Ernst Cloos and Hans Cloos, "Pre-Cambrian Structure of the Beartooth, the Bighorn, and the Black Hills Uplifts, and Its Coincidence with Tertiary Uplifting," *Proc. Geol. Soc. Amer. for 1933* (1934), p. 56.

Investigation of this problem has been carried further in the Beartooth Range by Lammers,[35] who has found that the borders of the Beartooth Plateau block, which forms the central and southeastern parts of the present uplift, coincide with the migmatitic border zones of the Beartooth batholith of pre-Cambrian age. "Laramide faults in the center of the range follow late pre-Cambrian shatter zones. . . . . In the area investigated no important Laramide structure was observed that was not closely related to an older structure in the basement complex."

Evidence pointing in the same direction has been presented by Blackstone, who has noted a close correlation between the attitude of the schistosity in the zone of pre-Cambrian schist exposed on the east margin of the southeast Pryor block and the position of the plane of the Dryhead fault which borders the block.[36]

The observed reflection and influence of pre-Cambrian structures in the later deformation is naturally attributable in part to local variations in resistance offered by the basement rocks and structures. Corresponding more or less closely to the pre-Cambrian structural pattern was inevitably one of varying resistance occasioned by the different strengths of different kinds of rock, by the pattern of faults or related zones of especial structural weakness, by differences in attitude of other structures, such as folds, schistosity, and intrusives, and by various combinations of these factors. Lines of least resistance in the basement complex were obviously guiding factors in localizing and determining the nature of the later deformation in this region where the overlying sedimentary cover was not excessively thick. On the principle of initial dip enunciated by Willis, a dome-like structure once started is a locus of weakness and will flex more with renewed stress, even if horizontally directed. Pre-Cambrian domes are likely to be rejuvenated or accentuated in later episodes of diastrophism.

CROSS SETS OF VERTICAL SHEAR PLANES

If one studies this portion of the Middle Rocky Mountain foreland area on the tectonic map of the United States now being prepared

[35] E. C. H. Lammers, "Pre-Cambrian Control of Laramide Structures in the Beartooth Range, Montana and Wyoming," *Bull. Geol. Soc. Amer.*, Vol. L (1939), p. 1918.

[36] Blackstone, *op. cit.*, p. 606.

by the National Research Council,[37] he is likely to be impressed by the prevalence of blocklike outlines. In many places the structure-contour lines turn nearly right-angle corners. These are where one of the east-west boundary lines already noted meets a north-south structural line. Such are the northeast corners of the two southern Pryor blocks, where the prolongation of the east-west Nye-Bowler lineament meets a nearly vertical north-south fault on the east side of each mountain block. Paralleling this lineament farther north is the Lake Basin line of echelon faults, at whose eastern extremity the contours bend abruptly south, forming two sides of a long rectangular block. In close parallelism still farther north are two more strong structural lines of the same general sort (one extending east from the southern flank of the Big Snowy Mountains to the south end of Porcupine Dome, and the other the Cat Creek anticline extending east to the north end of Porcupine Dome). Between them is another long, rectangular uplifted block with a nearly straight north-south eastern boundary. On the western side of these several blocks is a structural line trending about N. 25° E. which, in feebler fashion, extends the scarplike western face of the Beartooths for nearly 200 miles to the northwest corner of the Little Rocky Mountains.

The Beartooth Range is decidedly rectangular, the western and eastern halves comprising two rectangles, as we have already noted in considering Strips 1 and 2. On the western flank of the Black Hills the steep upflexure from the Powder River Basin shows an abrupt bend harmonizing in orientation with the Beartooth northeast corner and certain structures in the southern half of the Bighorns. The eastern limb of the Black Hills structure curves like the western side, through with less angularity, making the whole Laramide Black Hills uplift somewhat elbow-like in plan. Other similar cases are observed.

The widespread occurrence of this distinctive pattern indicates a common control on a large scale. The region has two sets of strongly developed structural lines nearly at right angles to each other. One set trends about 10° south of east; the other is nearly

[37] Or see C. E. Dobbin and C. E. Erdmann, *Structure Contour Map of the Montana Plains* (U.S. Geol. Survey, 1932).

north and south in some places, though trending somewhat east of north in other places. Both sets are oblique to the general northwest-southeast Rocky Mountain trend. Some of these lines are marked on the earth's surface by faults; rather more generally they are sharp upflexures between basins and uplifts; in some places, however, the Laramide vertical movement was not great. Many of them doubtless consist of lines of fracturing in the basement and sharp monoclinal folding in the sedimentary cover above. Their prevalence and the strength of their development are the striking features. What do they signify?

Application of the strain ellipsoid gives us a picture. In a short note a dozen years ago, the ellipsoid was applied, with two alternative orientations, to the structures of the Appalachians.[38] The first alternative was to place the longest axis of the ellipsoid of strain in a horizontal position extending northeast and southwest in line with the Appalachian ridges. As shown in Figure 2 of that article, the planes of maximum shear under compressive stress would in this case stand in a vertical position and would strike approximately 45° to the northeast-southwest elongation. But this orientation does not fit the Appalachian case. The dominant faults there do not cut obliquely across the long Appalachian folds in two sets nearly at right angles to each other, nor do they have approximately vertical dips. Instead they are typical reverse faults trending with the folds and dipping on the average between 35° and 40° southeast. This brings us to the second alternative, namely, that the greatest elongation (represented by the longest axis of the strain ellipsoid) was vertical and not northeast-southwest in alignment with the range. The shortening (least axis) being unquestionably northwest-southeast, the intermediate axis must lie northeast-southwest along the range. As thrust faulting normally occurs in, or close to, the plane of the intermediate axis and strongly inclined to the other two axes, this orientation fits the Appalachian case. Long, linear folds developed normal to the compression by upward relief, and there were only minor components of movement parallel to the intermediate axis or to the surface trend of the folded range.

[38] R. T. Chamberlin, "The Strain Ellipsoid and Appalachian Structures," *Jour. Geol.*, Vol. XXXVI (1928), pp. 85–90.

On the other hand, it is apparent that the first alternative, which did not fit the Appalachian deformation, does give the block pattern which we have just noted in the Middle Rocky Mountain foreland. Our two distinct sets of prominent lines of faulting, in some places observed and in other places assumed to exist beneath sharp monoclinal flexures and lines of secondary echelon faulting, are nearly at right angles to each other, and each is inclined about 45° to the Rocky Mountain trend. Also, where the primary faults are observed, they are very steep. For Laramide compression to produce fractures thus oriented, the strain ellipsoid would say that there must have been some horizontal elongation, as well as horizontal shortening, in this portion of the Rocky Mountains. The shortening having been northeast-southwest normal to the main mountain trend, this elongation would be northwest-southeast in line with that trend. The ellipsoid would say, furthermore, that the nearly vertical faults bordering the rectangular blocks are planes of shear under the assumed compressive stress, and that considerable components of the shearing were horizontal parallel to the strikes of the faults. The long narrow belts of echelon faults, which are best explained as tears in the sedimentary cover over lines of differential horizontal shifting in the basement, give us tangible evidence of horizontal displacement along these vertical shear planes.

Evidence seems to indicate that many and possibly most of these lines of faulting are much older than the Laramide deformation. If so, they would predispose the area to movements along the same lines of weakness in Laramide times, and thus be an important factor in determining the type of yielding in the later deformation. One may raise two questions: To what extent did pre-existing lines of fracturing determine the nature of the Laramide movements, and to what extent did the Laramide deformation contribute the lines now observed? Whatever the answers may be, there was Laramide horizontal shearing along these lines, as well as the very obvious differential vertical movement between blocks outlined by these faults.

In Montana one observes evidence of much more differential horizontal movement along the east-west lines than along those of the north-south set. Along each of the strong east-west lines (Nye-

Bowler, Lake Basin, Cat Creek anticline, etc.) between the Beartooth Range and the Little Rocky Mountains the south side of the zone of shearing has moved east, relative to the north side. As pointed out long ago in interpreting the Lake Basin strip of echelon faulting,[39] this differential movement is correlated with the orographic pattern and relative position of the mountain uplifts. South of the Lake Basin structural line the mountain uplifts (Beartooths, Bighorns, and Black Hills) extend several hundred miles farther east than the Rocky Mountain front north of it. The flexing and uplift of these ranges to the south involved more shortening there than occurred in the less-deformed plains lying due north of them. On the north the principal shortening occurred in the main Rocky Mountain chain much farther west.

In a word, the belt of Rocky Mountain shortening here shows "offsetting"—toward the west on the north, toward the east on the south. So great an offset in the belt of folding and crustal shortening necessarily required extensive crustal accommodation by east-west differential shifting of the two areas of differently distributed shortening. Relatively at least, the plains area to the north has shifted west with respect to the range and basin area to the south. This involved a strain strongly rotational in horizontal plan. The longest axis of this strain would be northwest-southeast; its shear planes would be vertical, one set approximately east-west, the other north-south. In this adjustment differential shifting of the sort noted above on the east-west lines was naturally more effective than movement on the north-south shear planes. The east-west shearing was recognized and interpreted first. Now, with further knowledge and recognition of the north-south shears, we can see how the two together have contributed the blocklike lines noted in certain portions of the tectonic pattern. This rotational strain affected an area from central Montana far into Wyoming.

COMPOSITE CONTROL

However the forces may have been applied, the resultant of compression, or the effective compression, operated along northeast-

[39] R. T. Chamberlin, "A Peculiar Belt of Oblique Faulting," *Jour. Geol.*, Vol. XXVII (1919), pp. 610–12.

southwest lines in the direction of greatest shortening. The observed structures of the area indicate that the yielding under this compression was of two types: (1) horizontal shearing along conjugate sets of near-vertical fractures, or shear zones, which has outlined to a varying degree various crustal blocks, and (2) vertical movements raising oblong ranges while basin areas sank. The horizontal shearing was somewhat peculiar, arising from special conditions; relief upward being, in general, easier than lateral relief, the dominant type of yielding was the rise of ranges and the sinking of basins more or less normal to the shortening.

Had the resistance of the materials not varied considerably from place to place, relief vertically would have been manifested in nearly parallel upfolds and downfolds. But as the Paleozoic and Mesozoic sedimentary cover was only moderately thick in this area, the underlying pre-Cambrian structures and competency differences were close enough to the surface to exert a strong influence both in localizing the horizontal shears and in locating and outlining the major uplifts by vertical easement. The special conditions which led to the rotational strain, above described, were presumably an inheritance from earlier deformations, and likewise, as we have seen, many of the details of the ranges rising under vertical yielding reflected earlier domes and other structures. The first contributed a tendency to block lines; the second, to ovate and oblong shapes; and both together predisposed certain areas to move as major units in the Laramide yielding. The combination interfered with regional parallelism. It is these modifying factors which have caused the departures of some of the ranges and basins from the general Rocky Mountain trend and have given the major structures of the region many of the details of their outlines.

In the vertical relief from the regional compression some units of the foreland under this composite control were forced up strongly as main ranges and tilted variously in the process; some small blocks, like the Pryors, were tipped up in trap-door fashion; other blocks, such as those north of the Beartooths and Bighorns, rose only moderately; while other areas of large size sank as basins. Between units, as they were crowded together in adjustment to the reduction in area, the sedimentary cover was strongly flexed. Sev-

eral of the raised mountain masses were a composite of smaller blocks, or segments, some of which were tipped up most on the northeast side, while others were elevated most on the southwest side in the general northeast-southwest shortening. Thus arose the main ranges of the foreland.

DEVELOPMENT OF THE LESSER FOLDS

The minor folds, formed both within the basins and on the ranges, differ in areal pattern from the features of larger relief outlined earlier. Instead of possessing notable irregularity of trend like the larger units, the minor anticlines in general are more regular in alignment. Most of them trend between N. 30° W. and N. 60° W., averaging around N. 45° W. The trends of these minor folds were influenced to some extent by the trends of the larger uplifts and basins, but in the main they are independent of them. This becomes strikingly apparent if one will but place the geologic map of western Wyoming on a table and, standing at the southeast corner, will sight northwest across the map. The minor folds show a parallelism running northwest-southeast throughout the whole region almost regardless of the orientation of the mountain ranges. What we see must be a regional folding superimposed on the irregular ranges and basins outlined under a different control. The pre-Cambrian basement pattern is supposed to have been a strong factor in outlining the main ranges and basins, but it was obviously much less a factor in determining the trends of the shallower minor folds.

The arrangement of these minor anticlines and related faults in both Wyoming and Colorado has often been called en echelon. But to the present writer, the echelon arrangement seems more apparent than actual, and the echelon concept itself seems to have obscured the true picture. These folds, by themselves, are hardly more en echelon than are short folds in general which pitch at their extremities and overlap one another. Their trends are regional, and they give the echelon picture only where they happen to be oblique to the long dimension of a mountain range, or elongate uplift, upon which they are superposed. This relationship is well shown in the southern Bighorn Range, which extends nearly due south and then curves toward the southwest, while the minor folds and related

thrust faults have the regional northwest-southeast trend obliquely across it. Similarly in Colorado, where the Front Range is essentially north and south, the regionally trending northwest-southeast minor folds and faults upon it produce the echelon pattern. It seems strange that this larger relationship has been so commonly overlooked.

FIG. 7.—Deformation of brittle paraffin under slight rotational strain. Fractures parallel to sides are shear cracks; the diagonal fractures are tension cracks. The shear fractures parallel to the arrows (applied force) are far more conspicuous than the set at right angles. From W. J. Mead.

The natural conclusion is that these minor folds of relatively parallel orientation were a direct result of the northeast-southwest regional shortening, influenced, so far as trends are concerned, only to a minor degree by the block shearing and pre-Cambrian structures which were such important determinants of the main ranges. The genesis of ranges was primarily a basement deformation; the minor folding and related faulting, on the other hand, were presumably relatively shallow, affecting chiefly the sedimentary cover and

reaching no great depth in the pre-Cambrian. Near the surface, upward relief was easiest, and the strata were compressed into many small folds normal to the shortening.

To illustrate the relation and the differences between the block shearing and the minor folds, we may again use Mead's experiments. In the case of Figure 7 his sheet of rubber was thinly coated with paraffin made brittle by cooling. This thin paraffin coating deformed very closely with the underlying rubber, and its shortening and chief elongation were both horizontal. It formed a network of fractures, including two cross sets of vertical shear cracks trending at about

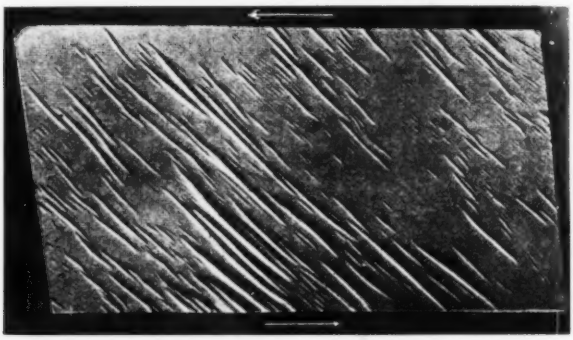

FIG. 8.—Deformation of plastic wax under slight rotational strain. Reduction in area in this case caused folding normal to the resultant of compression. From W. J. Mead.

45° to the direction of shortening. They represent our block pattern. In the case of Figure 8 the rubber sheet was coated with a layer of plastic wax covered with tinfoil, which supplied a small amount of competency. The plastic wax did not follow the deformation of the rubber sheet so closely as did the thin coating of paraffin, but found easier accommodation upward. It formed a succession of folds at right angles to the greatest shortening with but little corresponding folding in the underlying rubber. They are like our minor folds. Both types of deformation, however, did combine in the basement to produce the main ranges.[40]

[40] But although rotational strain, as illustrated by Fig. 8, was a factor in the development of the ranges, essentially nonrotational northeast-southwest compression may still have been the more general underlying cause of the foreland deformation.

Although the minor folds display an approach to parallelism over the whole deformed area, nevertheless within the area under consideration they have the same direction of asymmetry as the mountain segments on which or near which they formed. As the mountain segments were moved and tilted at the outset, so the near-by minor folds in general were turned later. In seeking the explanation of these strips of unidirectional asymmetry which involves the smaller folds as well as the larger mountain segments, and whose direction is reversed from strip to strip, three varieties of the operating process must be considered.

1. The segmentation of the ranges, as we have seen, was in some instances along east-west shear planes and in other instances, by tension cracking, transverse to the ranges as they were uplifted in the crustal shortening. We may suppose that the differential movement between segments of the Beartooths and Bighorns was extended well beyond the limits of those ranges, outlining the described strips of range and basin which, in the necessary adjustments to fit the space requirements, shifted alternatingly in opposite directions. With push in one direction dominating through a given strip, it would be natural for contemporaneously forming folds, large and small, to be harmonious throughout that strip. We may recall Leith's dictum: "As the little drag fold is inclined, so is the next larger unit of structure."

2. The unequally uplifted mountain segment having risen early, the direction of its tilting may determine the direction of inclination of the subordinate minor anticlines developing subsequently in the basins on either side. Push from a basin up the long gentle slope of a mountain segment will naturally turn the superficial anticlines so that their steeper limbs face the mountains, whereas the basin below the short, steep side of the mountain block might be expected to underthrust and cause backfolding of such anticlines as formed, though, as we have noted, relatively few anticlines did develop in this situation.

3. In bending strata into a syncline, adjustments between the beds requires the upper layers on the flanks to shift outward from the trough relative to the layers below, as may be seen by flexing a deck of cards. As Thom has pointed out, one might, therefore,

naturally expect the minor anticlines in these horizontally compressed basins to be turned outward from the middle of each basin. This process Thom has called "basin mechanics." But, instead of having this relation as a general characteristic, the anticlines in some strips have the same direction of inclination extending completely across both basins and ranges. The control in these cases therefore extended beyond the reach of any single basin, but it may perhaps have resulted from combination or competition of shove from several basins. On either side of each range is a basin, and shove from one basin may have dominated one segment of the range and of the basin beyond, whereas shove from the basin on the opposite side of the range may have been dominant in adjoining segments.

These three variations of the same general process may all have contributed to the observed results. On the gentler slope of a mountain segment all three would have co-operated to turn the smaller folds in the same direction as the larger uplift. Along the margin of the basin below the steeper slope of the mountain segment they would work more to cross purposes. In general, anticlines are poorly developed here. One does, however, find anticlines with conflicting directions of turning and thrusting in the Bighorn Basin at no great distance west of the steep southwest flank of the northern Bighorn segment (Strip 5). These are the Sheep Mountain folds, and the boundary between Strip 5 (southwest turning with the northern Bighorn segment) and Strip 2 (northeast turning) is drawn just east of them. But until more study has been devoted to this area, it is not easy to decide whether the Sheep Mountain deformation is properly regarded as a southern extension and feebler manifestation of the movement which uplifted the Pryor Mountains, or whether its peculiarities should be attributed primarily to outward shove from the Bighorn Basin, on the principle of basin mechanics, in competition with opposite shove from the northern Bighorn Mountain segment.

CONCLUSION

Finally in conclusion, the structural development of this portion of the Middle Rocky Mountains seems to have resulted, in the

large, from two main types of yielding to the compression which caused northeast-southwest crustal shortening. One type of yielding may be illustrated by a strain ellipsoid whose longest axis is northwest-southeast (the Rocky trend) and whose intermediate axis is vertical; its shear planes are nearly vertical and in two sets, one a little south of east and the other slightly east of north, both diagonal to the general Rocky trend. The other type of yielding is illustrated by a strain ellipsoid whose longest axis is vertical and whose intermediate axis is northwest-southeast in line with the mountains. The first was primarily shearing in the strong crystalline rocks of the basement, whose pre-Cambrian structural and rock pattern is thought to have played an important part in determining the details; the second, involving the basement strongly in the development of ranges and basins, but involving the sedimentary cover more particularly in the formation of the minor folds, occurred because of easy relief upward, especially near the surface.

It would seem that very early in the Laramide deformation there was differential movment within the crystalline basement in many places along both sets of vertical shear planes. To a considerable extent apparently this was renewed movement along old fault surfaces; presumably also, but to an unknown extent, it was slip along newly formed fractures. Magmas rose along some of the deep fractures as early in the Cretaceous as Judith River times. The blocklike lines in the sedimentary cover, so noticeable on the structure-contour map, especially in the northern part of the area, thus had their beginning. Perhaps no great amount of crustal shortening by diagonal shearing was required to outline the blocks. Basement blocks already outlined in pre-Laramide times could be elevated or depressed. With further crustal shortening accommodation by vertical yielding became increasingly important. Some blocks (as two of those in the stretch between the Beartooths and the Little Rocky Mountains) rose moderately, still retaining essentially their original outlines; other areas (like the Bighorns) were more strongly uplifted, displaying both the Rocky Mountain trend and the diagonal lines of block shearing; other areas were depressed as basins. Along the borders of uplifts the Paleozoic and Mesozoic sedimentary strata flexed in monoclines, many of which later split in faults. Perhaps

beginning fairly early, but certainly becoming more important later with increasing shortening, the superficial rocks folded into the widespread minor anticlines and synclines. Least resistance was upward, and these folds took a regional trend normal to the shortening. Except locally, pre-existing basement structures and the development of the main ranges and basins had relatively little influence on their trends, though they did determine, directly or indirectly, the direction of their asymmetry.

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FLUTING AND FACETING OF ROCK FRAGMENTS

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ABSTRACT

Rock surfaces are eroded by particle-laden currents of air and water so that under given conditions polishing, grooving, and fluting result. Polish is characteristic when the particles are fine and when abrasion predominates over solution in removal of material. Grooving and fluting in homogeneous rocks are caused by the more or less regular variation in abrasive intensity resulting from vortical movement of the fluid. When the vortices are shed from points of separation of flow from the rock surface, it seems that grooves elongate in the current direction may be cut. Under some conditions the vortices appear to be fixed in position and discontinuous flutes are cut. The stream flutes and wind flutes which have been observed are similar in form although the stream flutes are between fifteen and twenty times larger.

Under certain conditions fragments are faceted in particle-laden currents. Facets at high angles toward the current tend to be worn off uniformly and polished, while those inclined at low angles are cut by the most characteristic flutes and grooves.

Limestone fragments partially imbedded in alluvium may be faceted by solution during rain wash. Moisture driven by rain commonly produces solution effects consisting of an anastomosing and rugose system of ridges and rills on nearly horizontal surfaces and of deflected streamlike rills on steep-side and lee surfaces.

In erosion of limestone fragments by particle-laden air currents abrasion is dominant over solution. Rilled surfaces tend to be smoothed and polished. However, it is believed that all the various types of polish, fluting, grooving, faceting, and rilling may develop essentially simultaneously in one locality, presumably during the time span of a few tens of years The predominance of one feature or another depends upon the idiosyncrasies of the immediate neighborhood of the limestone fragment.

INTRODUCTION

The erosion of immobile rock surfaces in currents of air or water has often been noted by geologists. The cutting power of these currents is implemented by fine-rock particles in suspension, and the effective process has therefore been described as abrasion. Continued operation on stones produces fluting, faceting, and polishing. Solution plays a role whose importance is difficult to analyze because of the masking or dominating effect of abrasion, the more soluble rocks likewise being softer. Where there is no directed current, exposed soluble surfaces are irregularly etched and stones protruding from soil are faceted by rain wash.

The two processes, solution and abrasion, together or separately, produce analogous grooving and fluting in the surfaces of isolated stones, which may or may not be combined with faceting. The mechanics involved in a directed current are those governing turbulent

and vortical flow. In rain wash, water films are sometimes so thin and velocities so low that the flow is essentially laminar.

In streams with small suspended load and low velocity, channel cutting is accomplished by the abrasive action of rolling and sliding tools and by solution. The channel form tends to be smooth and regular. With high velocity and small suspended load the vortical movement of the water causes pothole erosion. Under certain conditions elongate depressions may be cut in the channel in the direction of the current. When the depressions are long and continuous, they may advantageously be termed "grooves." Under conditions of high velocity and large suspended load a variety of cut forms are developed on the channel itself and on the larger tools—not only potholes and grooves but also short and discontinuous depressions which are herein defined as flutes. The use of "flute" in this restricted sense geologically in distinction from "groove" is not incompatible with some specialized established uses, i.e., "in decorative art, a concave depression relatively long and of any form, the sides not necessarily parallel" (*Century Dictionary*). As the result of removal of material from the upstream face of boulders during fluting, facets are developed.

Faceting of stones occurs under subaerial conditions through solution of exposed protuberances or through abrasion by wind-borne sand, this process sometimes being supplemented by solution by wind-blown rain. In wind-faceting, wind-fluting may occur, the form of the flute being similar to that developed in a stream of water because both result from vortical movement of the particle-laden fluids.

SCULPTURE OF ROCK FRAGMENTS BY STREAM EROSION

STREAM-FLUTING

Stream-grooving and potholing on bedrock and boulders are common features of high-gradient streams. Stream-fluting, however, occurs most prominently in a stream of high velocity that also has a high content of suspended matter. An excellent example of stream-fluting in the Grand Canyon is figured and described by G. K. Gilbert:

> Nothing can resist the incessant impact of the fine siliceous particles, and the whole river bottom, including both solid rock and boulders, bears indisputable testimony to the mightiness of their work. Every exposed surface is polished

at least, and the most salient faces are deeply and beautifully carved. Plate X gives an example of this sculpture, in which the material wrought is a homogeneous, fine-grained limestone . . . . and the curved face lay nearly horizontal. . . . . The concave facets, of which the surface is composed, appear to be of the nature of paraboloids of revolution, the apices of which are turned upstream.[1]

FIG. 1.—Stream-fluted limestone boulder at 119 Mile Rapids, Grand Canyon, Arizona. The current moved from left to right. Note the elongation of the flutes in the direction of flow and the sharp overhanging lip frequently occurring on the upstream margin.

Lugeon has described stream-fluting in the Yadkin River of North Carolina.[2] He emphasized the analogy between this type of fluviatile erosion and eolian erosion.

Stream-fluting and the reduction of both bedrock and boulders in the Colorado River have later been described in more detail.[3] Although the flutes are most commonly found on limestone boulders

[1] "Report on the Geology of Portions of Nevada, Utah, California, and Arizona," *U.S. Geog. Surv. West of the 100th Meridian*, Vol. III (1875), p. 72.

[2] M. Lugeon, "Le Striage du lit fluvial," *Annales de geographie*, Vol. XXIV (1915), pp. 385–93.

[3] John H. Maxson and Ian Campbell, "Stream Fluting and Stream Erosion," *Jour. Geol.*, Vol. XLIII, No. 7 (1935), pp. 729–44.

in the deltas of the tributary streams, they also have been found on both boulders and bedrock of granite, pegmatite, schist, and basalt in the main river channel. Their occurrence on the insoluble rocks suggests that abrasion by water-borne silt particles has played a predominant role and that relative hardness, not solubility, determines the frequency and extent of development. It may be supposed,

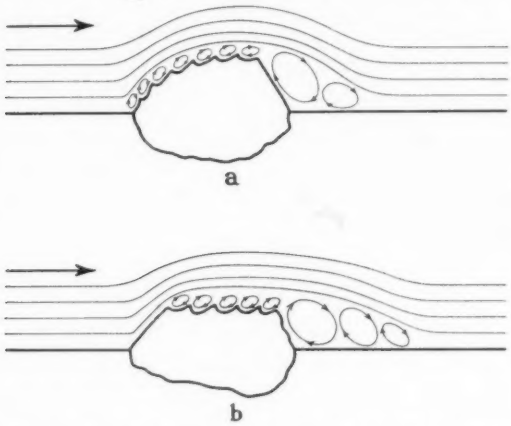

FIG. 2.—(*a*) Flow about an obstacle with upstream face at low angle to current direction showing vortices which might be developed thereon by skin friction. Pattern is intended primarily to represent water flow. (*b*) Flow about an obstacle with steep upstream face and point of separation with generation of vortices as a wake phenomenon. Under most conditions the vortices break away and move downstream. However, the erosion of flutes testifies to some stability under the natural conditions encountered. The vortex pattern is believed possible under proper conditions in either air or water. (Neither sketch is drawn to scale.)

however, that solution plays some part in the removal of material from limestone boulders.

The form of the flutes is determined by vortices formed by turbulent flow along large more or less flat surfaces and by disturbance of flow about irregularities of all types. The vortex axes tend to be subparallel to the water-rock boundary under conditions of uniform flow as along relatively flat bedrock surfaces or about the boulder section transverse to the general current as suggested in Figures 1 and 2. On the upstream area of boulders or other obstructions the

axes may be inclined at higher angles to the rock surfaces. Large vortices with slower rotation occur below the obstruction and are relatively ineffective in abrading the downstream surface. Examination of fluted surfaces, such as that shown in Figure 1, suggests that vortices of varying axial angle with the rock surface are present. Some flutes have spiral markings, others longitudinal markings, and still others are smooth throughout, indicating difference in the position of eddy rotation. Many of the flutes appear to be mutually interfering and encroaching. The extremely complicated and variable vortex pattern is probably due to change in river velocities passing from one flood stage to another.

STREAM-FACETING

The result of the differential removal of material from the upstream face of a boulder is faceting, the production of a rough face inclined at an angle of 50° or less to the direction of the impinging current. Faceting in streams is not a very common phenomenon because the condition of steady and rapid silt-laden stream flow not split by boulders is rare.

Stream-faceted pebbles in the bed of the Breede River, South Africa, have been described by Taljaard.[4] The faceted pebbles were observed to be

> restricted to the flanks of the main current where infrequent hollows or none occur, and where the depth of the water does not exceed 8 inches. . . . . Large boulders showing crude faceting on upstream faces are found at greater depths, but only where they directly face an unhampered floor- or bed-current, and are, in addition, encircled by sand.

This paper also notes that the better-shaped pebbles are those occupying the more stable positions either on a tenacious bottom or wedged in place.

A few examples of faceted limestone boulders have been observed in the Grand Canyon. Figure 3 shows a homogeneous, fine-textured limestone boulder from the upstream face of which considerable material has been removed. This face is rough but, in a general way, is inclined upstream at an average angle of about 50°. There

[4] M. S. Taljaard, "Note on the Occurrence of Faceted Pebbles as Products of Stream Flow," *Trans. Geol. Soc. South Africa*, Vol. XLII (1939), pp. 19–21.

is no evidence from the shape of the boulder as a whole that jointing is responsible for the face. The base is large, broadly rounded, and securely wedged among the other boulders of the delta. The boulders illustrated in Figure 4 occur on a bar and are less securely held in place. However, the upstream face of the limestone boulder is, neglecting surface irregularities and some curvature, inclined about

FIG. 3.—Limestone boulder with fluted and faceted upstream face. Arrow shows direction of flow of the Colorado River. Delta of Hermit Creek, Grand Canyon, Arizona. (Photo by Ian Campbell.)

40° upstream. The downstream face shows no scouring action and is intersected by the upstream face at an acute angle. The schist boulder in the foreground of Figure 4 shows strong evidence of abrasion on the upstream surfaces which are inclined about 45°and 30°, respectively. The schistosity is truncated by the 30° surface, and a few well-developed flutes are present. The bases of the boulders are large and irregular, and no suggestion of an imbricated or "shingle" structure is present.

RATE OF FACETING

No quantitative data on rate of faceting during stream-fluting have been obtained. General considerations, the probable stability of boulders during the seasonal floods, and the development of flutes on freshly exposed joint surfaces suggest that the process of abrasion in the Colorado River is very rapid. Flutes seem to have been formed

FIG. 4.—Fluted and faceted boulders; in front, boulder of quartz mica schist truncated across schistosity; in back, homogeneous limestone boulder. A six-inch scale rests on the limestone below the point of the arrow indicating current direction. Colorado River opposite the mouth of Trinity Creek, Grand Canyon, Arizona.

during such a short time as several years, and considerable reduction *in situ* or faceting probably occurs in a few tens of years.

SCULPTURE OF ROCK FRAGMENTS BY SUBAERIAL EROSION

WIND EROSION

The general features of wind erosion have long been known and described. The effects of wind-borne sand abrasion on large surfaces have been noted as pitted, etched, engraved, fluted, and grooved. Polish may be developed independently on smooth surfaces or may, in some cases, be formed on the fluted surfaces. Rock fragments are reduced in place and faceted. Illustrations of fluting on large

surfaces will be cited, and the occurrence of flutes on small faceted fragments will be discussed with reference to conditions of formation and mechanism of origin.

Blake described the forms produced by winds confined in San Gorgonio Pass between the San Bernardino and San Jacinto mountains of Southern California.

> The effects produced on the vertical surfaces of the rock exposed to the wind were, perhaps, the most curious and interesting, for here the hard minerals were left standing out in points, the softer feldspar being cut away on all sides. Masses of feldspar and quartz thus presented very rough and uneven surfaces.
>
> Where the feldspar was charged with small garnets, and was directly in front of the wind, a very peculiar result was produced; the garnets were left standing in relief, mounted on the ends of long pedicles of feldspar which had been protected from abrasion under the garnets while the surrounding parts were cut away. These pointed masses or needles of feldspar, tipped with garnets, stood out from the body of the rock in horizontal lines, pointing like jeweled fingers, in the direction of the prevailing wind.[5]

Gilbert likewise observed excellent examples of wind work.

> Blake and Newberry observed that in the Colorado Desert the pebbles were etched by drifting sand, and our examinations have detected the same phenomena in the Gila and Amargosa deserts, and other broad valleys. The most perfect work of this character that has fallen under my observation, is on a broad gravel mesa sloping gently toward the Colorado River, just below the mouth of the Virgin. The surface of compacted sand and gravel is hard, smooth, and even, and upon it are thickly strewn loose pebbles, shaped by the drifting sand. Hard, homogeneous stones, as quartzite and chalcedony, are rounded and highly polished, as though by collision in running water; crystalline rocks, as basalt and trachyte, are unevenly worn, the harder crystals maintaining prominences; the limestones are carved, with a network of vermicular grooves, into the most beautiful arabesque designs (Plate IX).[6]

The faceted cobbles shown in his Plate IX appear to have the typical vermiculated surface. They were formed by solution and modified by abrasion. Referring to a locality on Fortification Hill near Boulder Dam, Gilbert states:

> Boulders of basalt and trachyte, gradually weathering from the gravel of which the hill consists, are here subjected to a natural sand-blast of great

[5] W. P. Blake, in *Report of Explorations in California for Railroad Routes, etc.* ("Pacific Railroad Reports," Vol. V [Washington, D.C.: War Department, 1857]), Part II, p. 92.

[6] *Op. cit.*, p. 83.

power. The wind, confined by the hills, has locally but two directions—directions exactly opposed to each other—and these are conspicuously portrayed by the carving. In the larger of the basalt boulders represented in the illustration (Plate VIII), a point near the left margin faces the prevailing wind, and from it eroded grooves radiate over the convex surface; and the course of the blast, as it whirls around its edge, is finely shown on the stone at the right.[7]

These boulders split the wind. The sculpture described is analogous to the radiating flutes formed on boulders in a swift silt-laden stream. A pitted and fluted granite boulder near Garnet Station (Blake's occurrence) is figured by Blackwelder and ascribed to abrasion by wind-borne sand.[8]

An excellent example of wind-fluted limestone at Um Shersher, south of Kharga Oasis, Egypt, is figured by Hume.[9] These flutes are extremely elongate and presumably were cut under higher wind velocities than those described below from Death Valley.

Brunhes has described eolian potholes in rocks of the Nubian Desert.[10] These are 4–8 cm. in diameter and 6–10 cm. in depth. They occur most often on more or less vertical faces either facing the prevailing wind or facing away from it. Corrosion pits are also found in the same locality. Because some of the early pothole forms are of irregular orientation while better-developed forms are definitely oriented in the wind direction, Fleury believes that these developed as corrosion pits and were later modified by eolian corrasion.[11] Some of the potholes have been opened longitudinally and subjected to rectilinear friction.

Fleury described in detail the minor wind erosional forms he observed in Portugal. Grooved and fluted surfaces were attributed to two factors; one petrographic, which is concerned with the differential erosion of hard and soft parts of rocks; the other topographic, that is, direction of erosional processes by the accidental pattern

[7] *Ibid.*, p. 84.

[8] Eliot Blackwelder, "Cavernous Rock Surfaces of the Desert," *Amer. Jour. Sci.*, Vol. XVII (5th ser.; 1929), p. 396.

[9] W. F. Hume, "The Egyptian Wilderness," *Geog. Jour.*, Vol. LVIII (1921), p. 252.

[10] J. Brunhes, "Erosion tourbillonaire eolienne," *Mem. della Pontifica Accad. Romana dei nuovi lincei*, Vol. XXI (1903), pp. 129–48.

[11] Ernest Fleury, "Formes de desagregation et d'usure en Portugal," *Mem. Soc. Port. Sci. Nat., geol. ser.*, No. 1 (1919).

of the surface. In his interpretation the grooves are paths cut by means of rectilinear friction by the canalized sand between ridges. The ridges, themselves, would be miniature jardangs.

Boulders which split the wind and have elongate, radial grooves are illustrated by Fleury in his Figure 42. His fluted pebble (Fig. 52) closely resembles types found at the Death Valley locality described

FIG. 5.—View looking southwesterly over the sand dunes of Death Valley. The dune ripple marks indicate a prevailing southeast-northwest wind. The sands have been deposited about mesquite trees. In the foreground is the wash of Salt Creek, an ephemeral stream flowing from right to left. A number of properly oriented ventifacts were found in this wash.

below. Fleury notes that this type of grooving is found on horizontal or slightly inclined surfaces and is similar to stream-abraded grooves.

WIND-FLUTING IN DEATH VALLEY

The writer has observed wind-fluted and wind-faceted limestone cobbles and pebbles with polished quartzite cobbles and rilled limestone fragments in association near Stovepipe Wells and Salt Creek in the northern part of Death Valley, California. The ventifact lo-

cality is on the northeast side of Death Valley on the Boundary Canyon alluvial fan. The average annual rainfall at Furnace Creek Ranch, several miles south, is 1.4 inches. The general trend of the Death Valley graben is northwest-southeast, and a constriction in valley width just south of the ventifact locality is provided by Tucki Mountain of the Panamint Range. The wind tends to be confined to a northwest-southeast direction. The writer has personally observed a moderate wind from northwest to southeast in the evening and a strong wind from southeast to northwest in the morning during the month of December. Under some conditions the wind velocities may approximate those of a gale. Thus the ventifact locality is characterized by regular strong winds blowing from two opposing directions. With a low and intermittent rainfall, alluvial fragments may be relatively stable for several seasons. Hence, with an ample supply of fine debris available for tools, the locality is ideal for wind abrasion and wind-faceting. Three aspects of subaerial erosion of rock fragments are well exemplified. These are wind-fluting, wind-faceting, and rilling.

Freshly carved wind flutes are found only on limestone surfaces, horizontal or inclined at low angle in the direction of the prevailing wind. More or less modified and partly obliterated flutes may exist on other faces as the result of successive rotations. Quartzite faces and limestone faces inclined at high angle to the direction of the wind are smoothly polished. Most of the limestones which are fluted are homogeneous, and no suggestion of compositional or textural control for the location of hollows is present. It is believed, therefore, that wind flutes are cut beneath vortices by fine sand. The vortex, with its higher-velocity air current, is the locus of the more active abrasion.

The flutes are small, oval-shaped scallops elongate in the general wind direction. Typical examples are illustrated in Figures 6, 7, and 8. They are easily distinguished from stream flutes shown in Figure 1 by much smaller size (average length $\frac{1}{4}$ inch as compared with 4 inches), relatively much shallower depth, and less tendency to be overhung on the up-current end. Some flutes are elongate, and there are gradational types culminating in rather long, continuous grooves.

Inasmuch as the flutes are cut by impact and abrasion of sand grains beneath vortices, a fairly stable vortex pattern is indicated. The grooves, on the other hand, suggest vortices descending along the rock surface in the wind direction. For this reason the writer believes that the cutting of the grooves is dependent on the vortex pattern and not on canalization of sand between ridges, as suggested

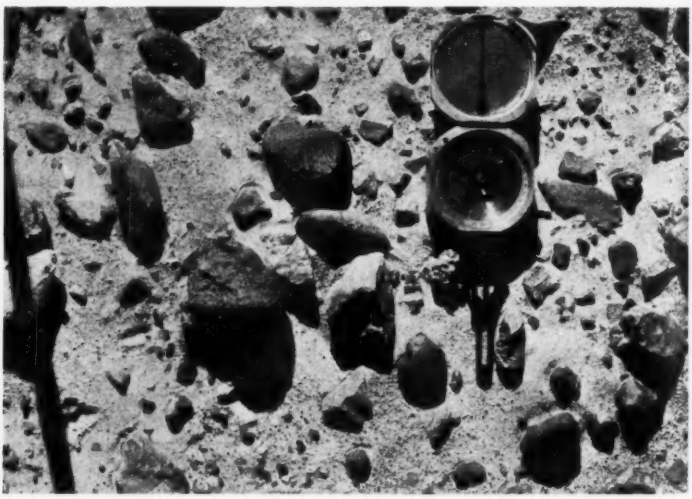

FIG. 6.—Undisturbed terrace surface showing limestone cobble with upper surface sharply wind fluted and faceted about 1 inch above and parallel to the ground level. Quartzite fragments show some rounding of corners and polish but are otherwise little eroded. The top of the photograph is south, and the prevailing wind blows from the southeast (upper left) to the northwest (lower right) in the general direction of the elongation of the flutes.

by Fleury. Once formed, the grooves may be modified by sweeping sand currents at air velocities below those critical in the generation of vortices.

MECHANICS OF FLUTE FORMATION

The flow pattern of fluids may, at higher velocities and under irregular channel or boundary conditions, become extremely complicated. Mathematical and empirical studies of turbulence under ideal conditions are still in part dependent on statistical analysis.

The theory of turbulent flow is the subject of intensive contemporary hydraulic and aerodynamic research. The flow of fluids in extremely complex natural channels under varying physical conditions,

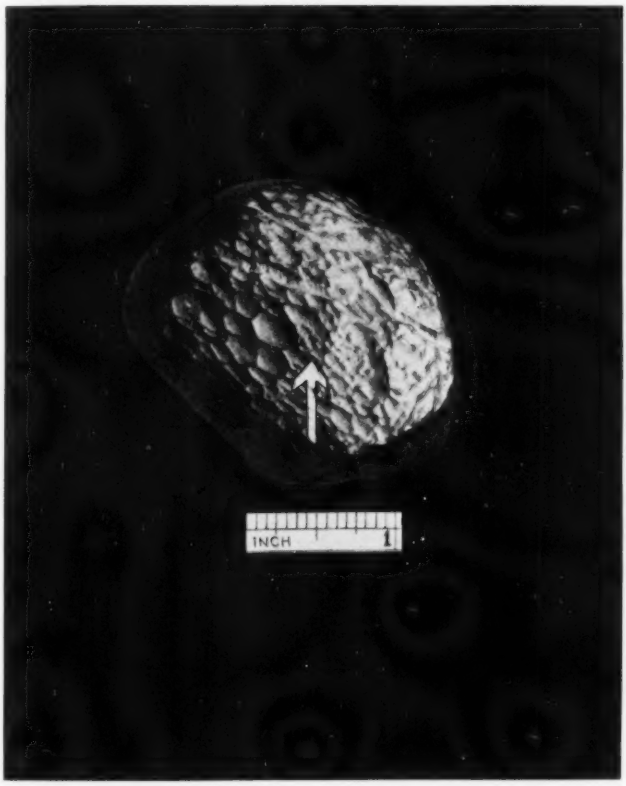

FIG. 7.—Strongly fluted horizontal facet. Arrow indicates north. Winds blow from lower right to upper left.

therefore, cannot as yet be analyzed rigorously. Nevertheless, consideration of the principles of fluid mechanics will assist in the understanding of the processes of erosion and rock forms resulting therefrom in currents of air and water.

There are three groups of variables which affect the flow of air

and water: (1) linear dimensions which describe the geometrical proportions, as, for example, in the case of an obstacle, the diameter d; (2) the dynamic characteristics of flow as the mean velocity V; and (3) the properties of the fluid with respect to weight, viscosity, surface tension, and elasticity. A generally used dimensionless parameter is the Reynolds number, $R = (Vd/\gamma)$, where γ is the kine-

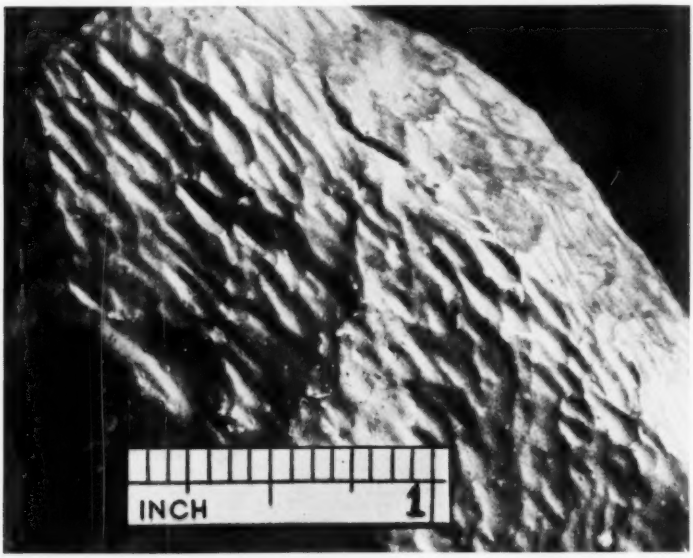

FIG. 8.—Wind-fluted limestone cobble surface, Death Valley. Flute shapes resemble those of stream flutes in Fig. 1, except that there is no overhanging lip. The wind swept from southeast (lower right) to northwest (upper left).

matic viscosity. The behavior of flow can be designated on a continuous scale of Reynolds numbers.

Streamline or laminar flow near a flat boundary involves the regular movement of a fluid without any local fluctuation in velocity, i.e., movement in layers without vortices wherein the layer at the boundary has a forward velocity of zero and those outward have successively higher forward velocities, the greatest increase taking place in the layers of fluid next to the boundary. In general, laminar

flow occurs only when the Reynolds number is very low because all disturbances are effectively damped by viscous action. Even considerable geometrical irregularities in boundary may not suffice to initiate vortices. However, with increasing values of R (increased velocity), disturbances are readily set up too great to be damped by viscous shear. At intermediate values of R along a nonangular boundary vorticity is confined to the boundary layer. Obstructions or curves in the confining channel (on the solid boundary) are causes of vorticity in some cases. In flow about a body where the curvature exceeds a certain critical value, vortices develop at what is known as the point of separation, i.e., where the streamlines leave the bounding surface and vortex sheets form the boundary layer. Vortices are then shed downstream, forming a wake. The vortices introduce a complex pattern of secondary motion in the fluid. In turbulent flow, masses of fluid are constantly changing position within the fluid body and are brought into regions of different velocity. The pattern of vorticity is usually illustrated by that developed about obstacles perpendicular to the flow. Such an obstacle generates vortex layers from each side known as vortex pairs which form the wake.

The mechanism of the motion may be explained in terms of the boundary layer and vortex layers. Separating the fluid in the wake from the main flow are the vortex layers which left the surface at the positions of separation. At moderately small Reynolds numbers the vortex layers, originating on the two sides of the body, come together farther downstream. Vorticity diffuses from the layers into the main body of the fluid, but is also generated in the boundary layer and added to the vortex pair. At small Reynolds numbers a state of equilibrium is set up between the rates of generation and the diffusion of vorticity, and the vortex pair is stationary relative to the solid body. As the Reynolds number is raised the rate of generation of vorticity increases; in order to counteract this the vortex layers become longer in the direction of flow, so as to provide a larger area from which vorticity can diffuse; at the same time the strength of the individual vortices in the vortex pair increases. The vortex layers are unstable; at some limiting Reynolds number they break away from the solid body and the vortices travel downstream.[12]

With increase to high Reynolds numbers the intensity of the vortices becomes greater, and they are more elongate in the direction

[12] *Modern Development in Fluid Mechanics*, ed. S. Goldstein (London: Oxford University Press, 1938), p. 552.

of flow. Likewise the higher the Reynolds number, the closer the vortices are formed to the point of separation.

Application of the foregoing principles to the practical cases of stream-fluting and wind-fluting is beset with difficulties. In the present state of our knowledge no rigorous application may be made. On any given surface in nature many irregularities of form may exist, and the velocity of air or of water flow is subject to irregular variation. However, by making certain assumptions, we may compute the Reynolds numbers for air under possible conditions at the Death Valley ventifact locality and for water in the Colorado River.

	Kinematic viscosity of air at 59° F at sea-level (effect of suspended matter neglected)	$\gamma = \frac{1}{6380}$
$R = \frac{Vd}{\gamma}$	Height of fragment	$d = 1$ inch
	Velocity (30 miles per hr.)	$V = 44$ ft./sec.
$R = 23{,}391$		
	Kinematic viscosity of water at 50° F (effect of suspended matter neglected)	$\gamma = \frac{1}{70800}$
	Height of fragment	$d = 1$ foot
	Velocity (7.5 miles per hour)	$V = 11$ ft./sec.
$R = 778{,}800$		

A Reynolds number of 2×10^4 is high and 7×10^5 is very high. In both cases the flow is turbulent. It has been noted that the size of stream flutes on midstream boulders is on the average sixteen times greater than wind flutes. Stream flutes on some channel walls in regions of lower stream velocity are noticeably smaller than those found on midstream boulder deltas. Thus it would appear that the size of the vortices is related to the Reynolds number. The lack of correspondence in magnitudes (flutes twenty times greater produced with Reynolds number one hundred times greater) may be due to the approximate nature of the Reynolds numbers used. For the computation in the Colorado channel a value of velocity, V, was taken from the mean velocity in the total cross section during flood, whereas the velocity decreases greatly toward the channel walls. The Reynolds number in the vicinity of submerged boulders must

be much lower than the figure given. Variation in the nature of the suspended load introduces another unknown factor.

The distribution of stream flutes on channel surfaces and on the upstream faces of boulders suggests that vortices induced by skin friction are responsible. Minor surface irregularities (roughness) doubtless complicate the pattern. The flutes on boulder peripheries across the direction of flow (i.e., rock surfaces downstream from a shoulder and upstream from another abrupt curve) appear to show some regularity in spacing and size as though formed in vortex trains below points of separation. Formation as a wake phenomenon is also suggested by the confined occurrence of flutes on the periphery of a lava boulder found near Lava Falls below the first points of separation. No flutes are formed on the downstream face, owing to diffusion of energy from vortex layers into the stream.

The regular distribution of stream flutes is noteworthy, inasmuch as vortices are believed to be shed from the point of separation above critical Reynolds numbers. Presumably the flow in the Colorado River must have a value conducive to stability of position. Once formed, the flutes may themselves have the effect of fixing vortices. One might expect the vortices to show some change in distribution and size in response to varying conditions of river flow. Such a change doubtless occurs. A theoretical distribution of flow lines about a boulder immersed in a stream of water is shown in Figure 2, *a*.

Wind-fluting has been observed so far only on horizontal facets and on surfaces inclined at low angles to the wind. The sides show no signs of concurrent fluting, although older fluting cut in a former position may be present. In the fluted fragments studied there has been a shoulder upwind from the flutes. This suggests that the wind flutes are a wake phenomenon formed in a vortex train below a line of separation. This suggestion is utilized in showing flow lines of air about an obstacle in Figure 2, *b*. If the shoulder were rectilinear and oriented at right angles to the wind, it would be expected on theoretical grounds that the vortex sheets would erode parallel troughs simulating ripplemark elongated transverse to wind direction. The actual pattern is one of many individual flutes sometimes showing a rough transverse distribution.

In both air and water the differentiation of vortex sheets into

regions of greater and less velocity of rotation, sometimes a little offset from one another, seems difficult to explain. However, the rough periodicity of occurrence (shown to greater or less degree—less in Fig. 1) would indicate that the individual flutes are formed in a regular system of vorticity. The grooves sometimes observed on the Death Valley ventifacts but not on the Colorado River boulders appear to be formed when the vortices are not stable but descend downwind from the shoulder of separation.

Unlike the stream flutes formed under flood conditions in a confined channel wherein consistently high velocities are maintained, the wind flutes are formed in fluid not confined by definite lateral boundaries and in which great and rapid fluctuations in velocity may take place. It is difficult to understand how vortices, sufficiently stable to erode flutes in solid rock, could be formed under such conditions. Possibly the fluting proceeds only in the upper range of velocities, and a set of flutes once formed may influence location of vortices under moderately variable velocities.

Questions regarding the vortex and flute pattern and its stability must await further experimental work. Additional discussion at this time is not justified by the laboratory and field data available.

WIND-FACETING

The subject of wind-faceting has been reviewed by Bryan,[13] a nomenclature set up, and a bibliography of 258 titles presented. Bryan has briefly summarized the development of the theory.

The formation of facets on stones by the wear of wind-driven sand was correctly analyzed by Woodworth in 1894. He held that the effectiveness of the sand blast as shown by Tilghman is greatest on surfaces between 30° and 60° from the horizontal. A facet is formed at right angles to the wind and once attaining an angle of 30° is thereafter only slowly modified. Other facets are produced in shifting or overturn of stones. The formation of stones by two contrary winds as held by Travers, Enys and Mickwitz is entirely feasible but only one wind is required. Bather (1900) in his admirable review follows Woodworth, but continental geologists have usually followed Heim who believes that the wind splits on the edges (Kanter).[14]

[13] Kirk Bryan, "Wind-worn Stones or Ventifacts—a Discussion and Bibliography," *Rept. Committee on Sedimentation for 1929–30, Natl. Res. Coun. Circ. No. 98* (1931).

[14] "Reviews of Papers on the Geomorphology of North America, 1933–35," *Zeit. für Geomorphology*, Vol. VIII (1935), p. 5.

Schoewe has shown experimentally that ridge-shaped ventifacts oriented at right angles to the wind can be formed by constant winds irrespective of the shape of the base of the model and that the rate of abrasion on faces sloping 30° is about one-third as great as on faces inclined at angles of 60°.[15] It was suggested by Schoewe that

FIG. 9.—Top view of a ridge-shaped limestone ventifact from Death Valley. In its original position the long axis was northeast-southwest perpendicular to the prevailing wind directions. Opposing facets inclined at 25° to the horizontal and fluted. The soil line is shown by sand grains lightly cemented to the cobble by lime.

in order to be faceted the ratio of height of the model to the height of the sand-laden current should not exceed 1:8. In a sand stream extending to 24 inches above the ground the largest fragment faceted would have a height of about 3 inches. Larger fragments would split the wind. This generalization is supported by observations in Death Valley, where the faceted fragments have a height of less than 3

[15] Walter H. Schoewe, "Experiments on the Formation of Wind-faceted Pebbles," *Amer. Jour. Sci.*, Vol. XXIV (5th ser.; 1932), pp. 111–34.

inches above the ground. Larger fragments are in many cases vermiculated, grooved, and polished, but they are not faceted.

The faceted fragments at the Death Valley locality are usually composed of limestone or highly calcareous shale. The most common type is ridge shaped, with a well-developed face to the southeast and a less well-developed face to the northwest. The ridge is

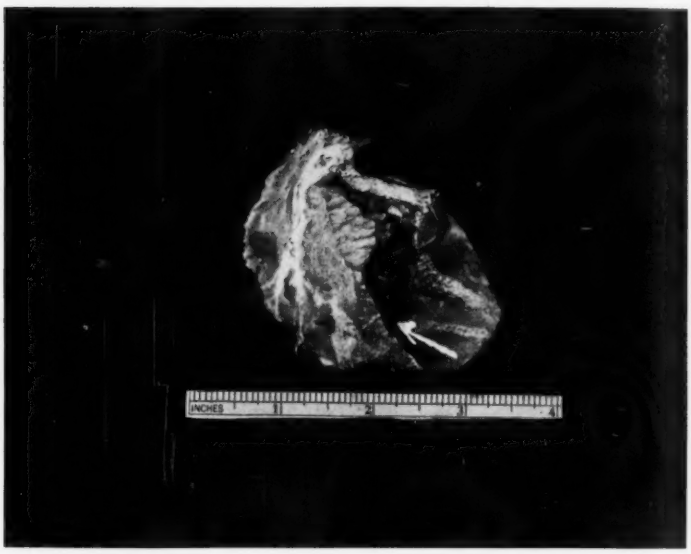

FIG. 10.—Faceted limestone cobble showing horizontal fluted facet and spurs facing principal wind direction.

oriented in a large majority of cases at right angles to the wind, checking with the observations of Wade in the Nubian Desert.[16] Wade noted between three and four hundred ventifacts, of which 78 per cent were set at approximately right angles to the wind and 22 per cent were set approximately parallel to the wind.

Figure 9 shows an example. The ridge was oriented at right angles to the direction of the wind. The cut faces are inclined at 25° to the

[16] A. Wade, "On the Formation of Dreikante in Desert Regions," *Geol. Mag.*, Vol. VII (1910), pp. 394–98.

horizontal and do not extend evenly to the soil level. The facets presumably formed intermittently under the action of opposite winds. Faint flutes are developed on both facets. The lightly ce-

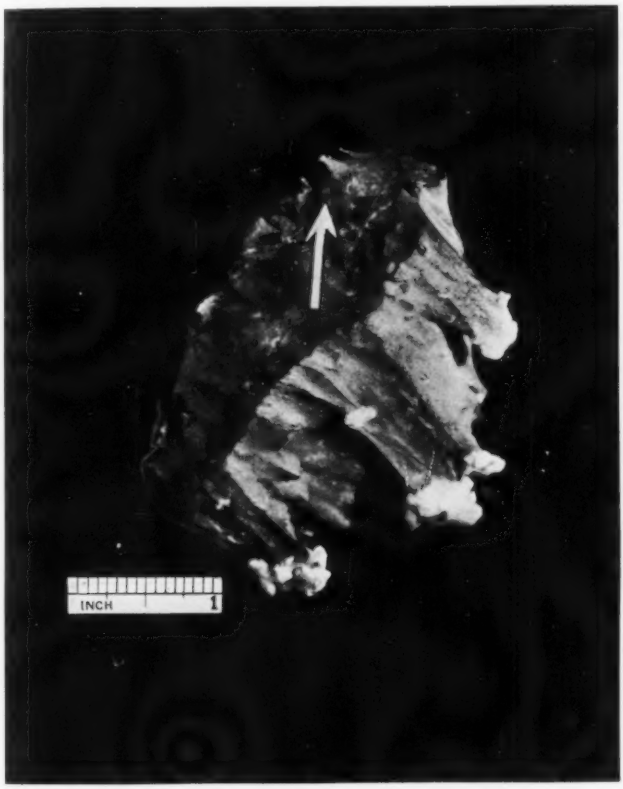

FIG. 11.—Etching of spurred facets on impure limestone. Spur tips supported by resistant silica nodules. Strongest of the opposing winds blew southeast to northwest (lower right to upper left).

mented sand adhering around the soil line offers the suggestion that to some degree solution may supplement the predominant abrasion in reducing the exposed surface. Other examples of ridge-shaped ventifacts are shown in Figures 10 and 11. In Figure 10 the oppo-

site facets are inclined 50° to the horizontal, and the ridge is truncated by a small horizontal, faintly fluted facet. Calcite veins have exercised little control on the sculpture, but minute siliceous bodies and veinlets have supported projections from the facets. Figure 11 shows the striking development of pedicles of limestone pointed by silica nodules facing the prevailing winds in a manner recalling Blake's description.

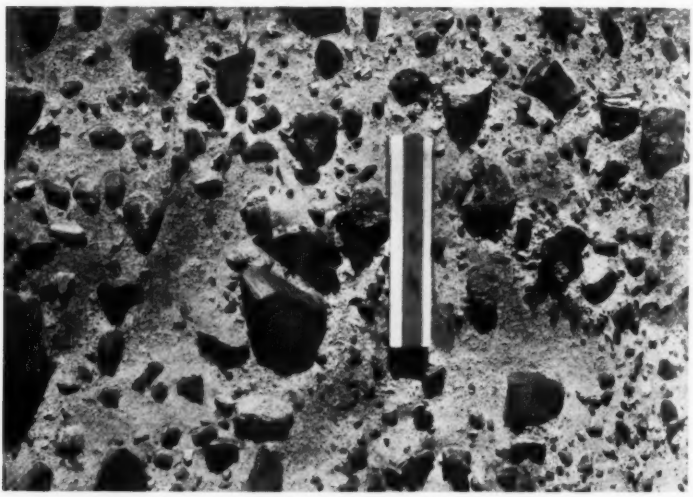

FIG. 12.—Typical litter on alluvial surface in the vicinity of the sand dunes, Death Valley. Note limestone fragment in lower left of center with top truncated parallel to and about 2 inches above the soil surface. In the upper right corner is another horizontally faceted pebble with fluted surface. The upper end of the 6-inch scale points south. Wind direction is from upper left to lower right.

Gradation in the facet angle with the horizontal from 50° (rare) through 25° (common) to 0° (rare) is noted. A few of the horizontal facets were reduced to, or almost to, the soil level, but many were found 1 or 2 inches above this level. The high-angle facets in homogeneous limestone were observed for the most part to be smooth and polished. The low-angle facets were observed to be fluted and polished. Figure 12 illustrates a limestone pebble with a horizontal facet carved with faint southeast-northwest flutes. Figure 6 shows

another example with well-defined, oriented flutes on the horizontal facet.

The rare occurrence of horizontal facets brings up the question of whether these may have been developed *in situ* or formed in some other position, either at a higher soil level since excavated or at an angle to the wind and later rotated. No cases of horizontal faceting are on record. Flutes, however, may be developed on horizontal surfaces. Areas of several square yards of a horizontal limestone bedding plane were found stream-fluted in the Marble Canyon of the Colorado River. A flat surface such as a joint plane, therefore, might conceivably be fluted by sand abrasion if near the ground level. However, some irregularity in the boundary between the air and the rock fragment may help to initiate the generation of vortices. Hence it seems difficult to eliminate the possibility of having the fluting proceed on a more or less horizontal surface above the general ground level. If the cutting of flutes occurs in this position, then at least some faceting likewise takes place in this position.

SOLUTION-FACETING

The importance of abrasion of rock surfaces by wind-borne sand in the cutting of facets is easily recognized, but the role of solution in facet-cutting is sometimes difficult to evaluate. Only in an environment where there is no strong and prevailing wind may the effects of solution alone be analyzed. It is, therefore, worth while to examine the conditions under which solution-faceting occurs and under which the features are formed.

The development of flat faces on portions of limestone fragments exposed above the land surface has been described by Udden[17] and by Bryan.[18] Udden ascribed the flattening of the pebbles to the solvent action of rain water. He considered that the flattening required a long time because the pebbles had been turned over many times. Bryan described in detail the various forms of faceted pebbles. The lime dissolved from the exposed surface was shown to be deposited as travertine on the pebble below the soil surface. The

[17] J. A. Udden, "The Flattening of Limestone Gravel Boulders by Solution," *Geol. Soc. Amer. Bull.*, Vol. XXV (1914), pp. 66–68.

[18] Kirk Bryan, "Solution-facetted Limestone Pebbles," *Amer. Jour. Sci.*, Vol. XVIII (5th ser.; 1929), pp. 193–208.

travertine, resisting solution more than the original limestone, was shown to form a rim during later stages of faceting. The exposed surface, at first hemispherical, was later reduced to a low surface with a prominence or boss near the center.

Solution-faceting of this general type has been observed by the author on the Coconino Plateau and on the Tonto Platform in the Grand Canyon district. On the Coconino Plateau many fragments of the dense, gray Kaibab limestone (Permian) are found to be

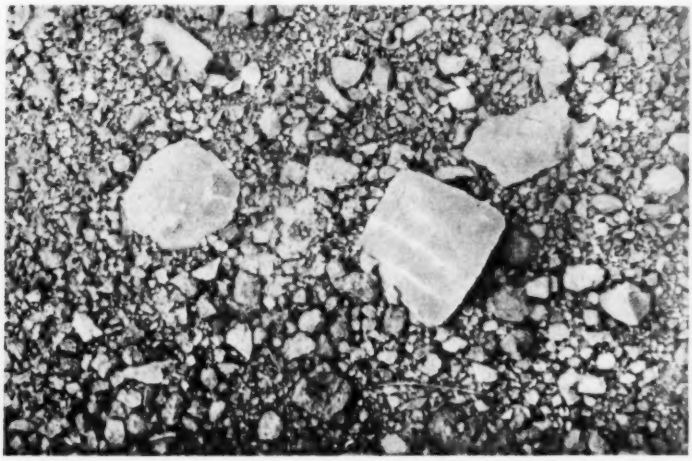

FIG. 13.—Faceted limestone fragments *in situ*, surrounded by small sandstone and shale fragments. The largest limestone facet is approximately 4 inches square.

faceted with rounded or hemispherical exposed surfaces. On the Tonto Platform fragments of dense, white Redwall limestone (Mississippian) are occasionally found faceted, and fragments of the soft, sandy Muav limestone (Cambrian) are faceted in large numbers. The various forms illustrated by Bryan are found. However, the geomorphic environment is not the same. Bryan's specimens occurred on a plain in Montana and on gravel terraces in New Mexico. Udden's occurrence was on Pleistocene terraces of the lower Rio Grande in Texas. The Tonto Platform cannot be regarded as old as or as stable as any of these surfaces. It is a sloping bench

cut in soft Bright Angel shales (Cambrian), controlled by the more resistant, cliff-forming, underlying Tapeats sandstone (Cambrian). The platform is drained by numerous small stream courses which are occupied intermittently during rainy seasons. The faceted fragments lie embedded in a gravelly soil (Fig. 13) on the interfluves. Some occur on slopes of 5°–10°, but on the steeper slopes disturbance by

FIG. 14.—Limestone fragment solution-faceted on upper and lower surfaces. Tonto Platform.

creep is rapid, and solution-fretted, but not faceted, fragments are more common.

As pointed out by Bryan, the central boss is more prominent on some pebbles than others, and his explanation that this provides a slope down which water may flow appears likely. The natural corollary of this is that a larger amount of water passes over the peripheral area of the face, thereby dissolving more limestone than at the center. This action would tend to maintain convexity. All the larger facets found on the Tonto Platform possessed some central convex-

ity. Only some smaller and possibly more porous fragments have nearly flat or actually concave facets. In these cases the margins of the face were supported by the rim, and removal by solution was more or less uniform over the whole surface or was actually greater in the center. Figure 14 shows a thin limestone fragment which has evidently been beveled on its lower surface, then turned over and beveled on its upper surface. Figure 15 shows a typical solution-

FIG. 15.—Solution-faceted limestone fragment showing characteristic central boss, travertine rim, and travertine deposited on lower surface. Tonto Platform.

faceted fragment with one face. About its irregular bottom is a heavy deposit of travertine.

Likewise recorded by Bryan is a variable amount of the secondary calcite in different occurrences. In some cases the lack of travertine may indicate an incipient stage of faceting. However, on the Tonto Platform the fragments lacking travertine were found to lie on the steeper slopes where one might expect the rain solution to be readily conducted away either down the surface or through the soil. Figure

16 shows such a fragment. Only a slight amount of secondary calcite has been deposited on the undersurface. On the exposed face several rills occur which follow fine fracture planes. Some of these etchings

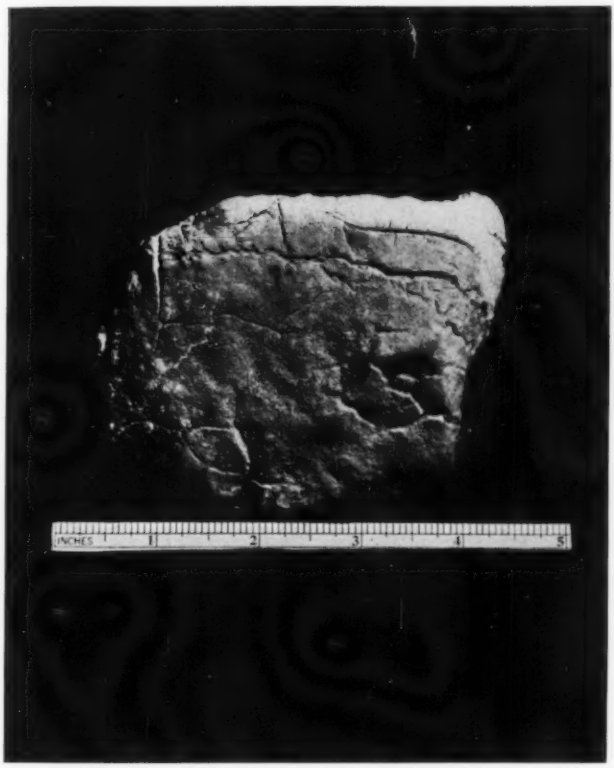

FIG. 16.—Upper surface of limestone fragment showing etching by rills controlled by fracture pattern.

of the structural pattern may be traced to the lower surface where, however, they are not so broadly opened. Figure 17 shows a fragment the exposed surface of which is fretted and pitted. The coating of secondary calcite is very thin. This fragment occurred on a steeper slope than the typical faceted forms. Although the lime was being

removed irregularly, it was, nevertheless, flattening the face. The pitting seems attributable in part at least to lack of homogeneity in the limestone.

In solution-faceting the agent is freely falling rain. In the Grand Canyon instance wind erosion does not seem important, judging from the experience of many weeks of field work on and about the Tonto Platform during various seasons. The consideration of this

FIG. 17.—Solution-fretted upper surface of limestone fragment from steep slope

case may be useful in evaluation of the role of solution in a region where wind erosion is dominant.

RATE OF SOLUTION-FACETING

The surface of the Tonto Platform on which solution-faceted limestone fragments are found is young geologically and subject to frequent disturbance. It suggests that solution-faceting may be relatively rapid in some cases and, therefore, does not in itself testify to the antiquity of an erosion surface. Bryan, in estimating age significance, stated:

> In default of local data an estimate of the rate at which solution of the pebbles takes place seems impossible, but we can be certain that the formation of a facet which involves the loss of an inch of limestone required hundreds of years. . . . . A surface paved by solution-facetted pebbles doubtless has, however, an antiquity of at least two or more thousands of years.[19]

The Tonto Platform has an arid climate, with considerably less than the 17 inches average annual rainfall of the south rim. But it is occasionally visited by local cloudbursts or heavy showers. Judging from observation through about five seasons, the writer believes that each part of the surface is visited by sufficiently heavy rains to disturb the surface and move fragments of cobble size within a few tens of years. On the basis of such physiographic evidence the writer believes that solution facets are formed in less than one hundred years on the Tonto Platform. It is likely that the limestone most frequently faceted here is considerably more porous and more rapidly soluble than in the cases discussed by Bryan. Care must therefore be exercised in ascribing antiquity to a surface containing solution-faceted limestone fragments.

COMBINATION OF SOLUTION WITH ABRASION

Limestone surfaces with patterns of arabesque design are found in different types of physiographic environment. Vermiculated cobbles found on a broad mesa below the mouth of the Virgin River in southern Nevada were figured by Gilbert and ascribed to carving by wind-blown sand.[20] Here the conditions are arid. Fleury has discussed in considerable detail arabesque limestones found in the limestone coastal areas of Portugal.[21] The phenomenon here is believed to be transitional between types found in desert regions and in the Alps. Fleury terms the feature "guillochage," literally "engine-turning," because of the complicated microsculptures. He discovered that the microsculptures do not pre-exist in the rock but are developed by erosional processes in some cases more or less influenced by petrographic factors. Corrasion by sandblast does not explain the features or their occurrence in Alpine regions. The fineness and the irregularity of the grooves in vermiculated types are

[19] *Ibid.*, pp. 207–8.

[20] *Op. cit.*, p. 83, Pl. IX.

[21] *Op. cit.*, pp. 119–25.

opposed to solution by water in simple runoff. Fleury describes the formation of vermiculations:

> In "guillochage" the water does not work as it does either in streams or in lapiesation; it avoids gravity, while it is agitated under the influence of wind currents which drive it forward from the point where it strikes the rocks. It is generally in the state of drops, and it is provided generally by the dew and by the condensation of vapor or of mists as well as from rain; its mass is very small and without true mechanical action. The solution is also directed at times by the control of the wind, by the details of the shape (or the rock), and by the petrographic factors, that is to say, by conditions narrowly dependent on one another, which are able to explain, directly or by their variations, all the forms of microsculptures, as, for example, the radiating position of the holes, the regularity of the vermiculations apparently directed by the slope, and, above all, the differentiation of the decorations on the same rock or on a single small surface.[22]

Fleury notes that, although solution is the principal process in erosion of the vermiculations, they may be modified by corrasion.

Figure 18 illustrates a stone of the type figured by Gilbert whereon the pattern is anastomosing. The cobble illustrated shows some variation in composition along bedding planes, yet the corrugations are not influenced thereby. The pattern reminds one of that produced by a film of water blowing over an automobile windshield during a light rain. The effect may be produced by concentrations of water, resulting from turbulence in the wind, with attendant solution along the surface. Once initiated, such a pattern might be perpetuated or accentuated by succeeding windy rainstorms. At the Death Valley locality rillensteine with streamlike pattern may be found. These are evidently formed, as suggested by Laudermilk and Woodford, by solution in streamlets flowing down the sides.[23] Some deflections of the streamlets, as the result of the wind, is shown in Figure 19. Laudermilk and Woodford also point out that the vermiculations and rills may be more or less completely smoothed out by sand abrasion. The concept that the rills are formed at one time and partially obliterated at a later time is suggested by Blackwelder.

[22] Translated by Kirk Bryan.

[23] J. D. Laudermilk and A. O. Woodford, "Concerning Rillensteine," *Amer. Jour. Sci.*, Vol. XXIII (5th ser.; 1932), pp. 135–54.

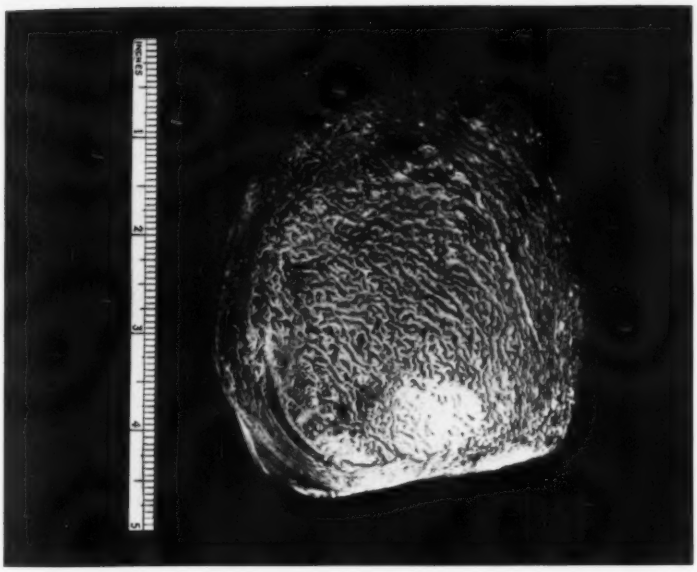

FIG. 18.—Solution pattern developed on smoothly rounded upper surface of a rather homogeneous limestone cobble. Death Valley, California.

FIG. 19.—Streamlike rillensteine with rills showing deflection from the top toward the left in response to the prevailing wind. Surface which was below soil unrilled. Death Valley, California.

Some of the crooked grooves which Eklaw and I have ascribed to solution, are limited to one or two sides of the pebbles. On the other side, which was more exposed to the sandblast, they have probably been erased. It seems to me

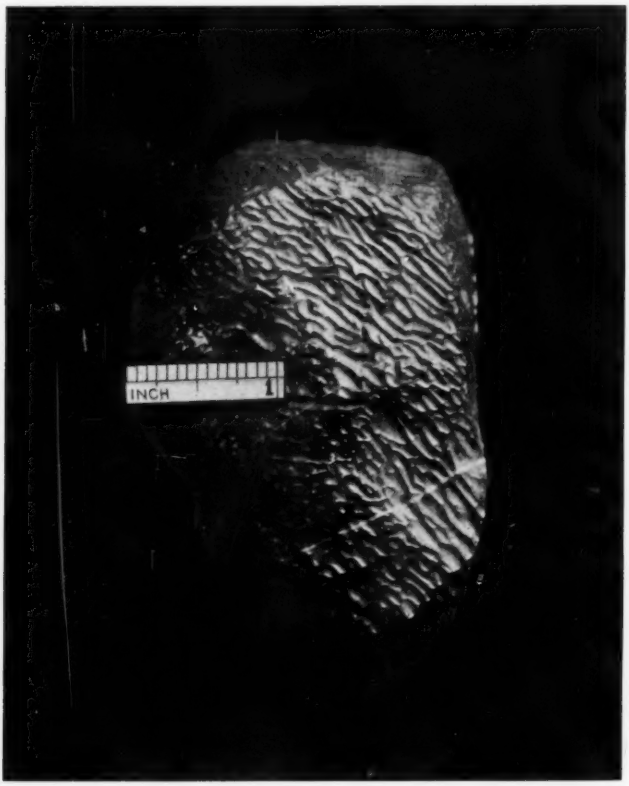

FIG. 20.—Cobble showing combination of fluting and anastomosing rilling on horizontal facet; truncation, deep fluting, and polishing on the windward face (lower right); and faint rilling and polishing on the lee.

possible that these stones preserve a record of climatic changes,—the rilled surface being made in somewhat moister epochs, and the sandblast facets being made in the dryer and perhaps windier times intervening.[24]

[24] Eliot Blackwelder, quoted by Laudermilk and Woodford, *op. cit.*, pp. 146–47.

It has been noted earlier in the discussion that the phenomena of faceting, polishing, fluting, and rilling are all found under similar conditions at the one locality in Death Valley. It is the opinion of the writer that separate epochs of solutional and abrasional activity are not required. The processes forming these features are all operative at the present time. The forms are often found in combination.

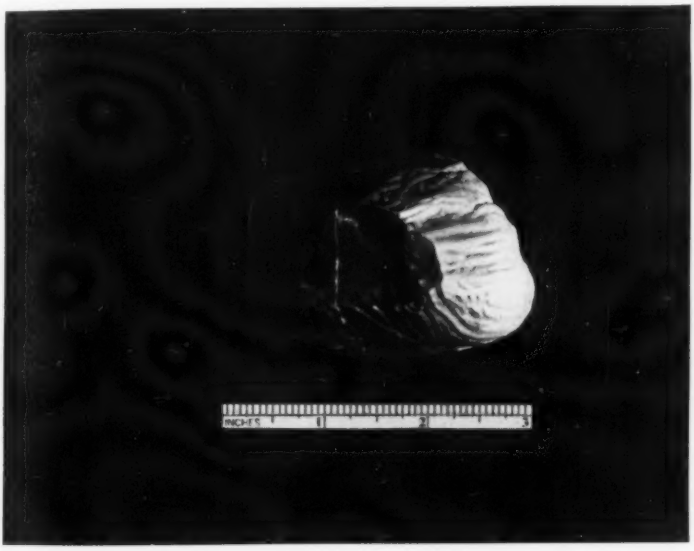

FIG. 21.—Top view of limestone cobble showing smoothly polished and cut windward face (*left*); fluted top or horizontal facet; and polished rills on lee side. Death Valley, California. Specimen presented by H. D. Curry.

Small limestone chips lying flush with the ground level resemble solution-faceted chips but have an irregular pattern of low ridges and hollows. Facets of ridge-shaped ventifacts disturbed in orientation have a polished rugose surface. Figure 20 illustrates a limestone cobble whose horizontal facet has a pattern of wind flutes in part and anastomosing rills in part, suggesting concurrent solution and abrasion of the surface. The windward facet is polished smoothly

for the most part but locally is fluted and is encroaching on earlier established flutes of the horizontal facet. This merely indicates that the direct sandblast erosion is much more rapid than the erosion of the flutes by vortices. Figure 7 shows a pebble on which well-marked wind flutes may be seen on the horizontal facet. The smoothly polished windward face has been reduced more than a

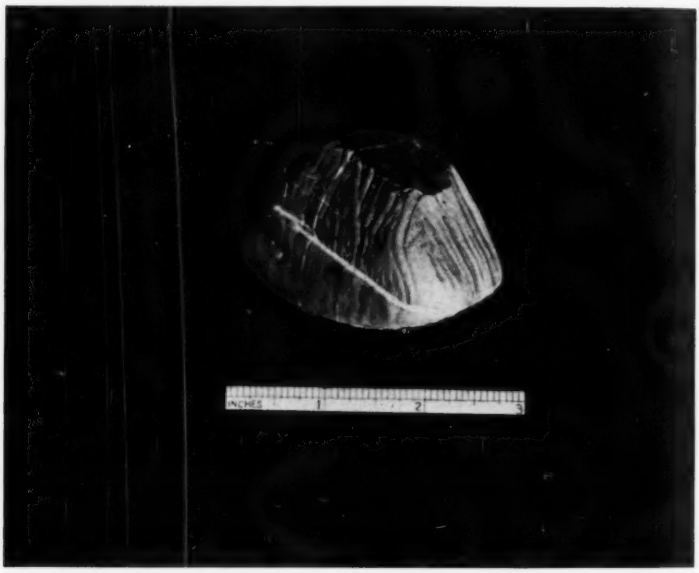

FIG. 22.—View of side and back of cobble shown in Fig. 20. Shallow fluting is visible on the horizontal facet. The broad and polished rills of the side appear to be deflected in the direction of wind flow, from left to right.

quarter of an inch as shown by a siliceous projection (hidden in the photograph). All features are shown on the cobble illustrated in Figures 21 and 22. There is a smoothly polished windward facet, a fluted horizontal facet, and extensive rilling down the sides deflected leeward and down the lee face. Some of the flutes are sources of rills. Although shallow and polished, it might be expected that they would be entirely obliterated by the rapid abrasion if they were

not occasionally renewed by solution. Thus it is shown that solution is an important subsidiary process in faceting of limestone fragments, even under conditions of extreme aridity.

ACKNOWLEDGMENTS.—The field work in the Grand Canyon district was carried out in conjunction with studies of the Archean rocks, a geological project, supported by the Carnegie Institution of Washington. The field work in the Death Valley region was supported by a grant from research funds by the California Institute of Technology. The writer wishes to thank Professor Kirk Bryan for many valuable suggestions and for criticism of the manuscript.

NEW MEASUREMENTS OF THE RIGIDITY OF ROCKS AT HIGH PRESSURE[1]

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Harvard University

ABSTRACT

New dynamical measurements of the velocity of shear waves and of the rigidity at high pressures are presented for a few igneous rocks, and for a group of sedimentary and metamorphic rocks from the Boston area. The results include the changes of velocity on heating to 150° C., at 4,000 kg/cm² and at 9,000 kg/cm² for the igneous rocks, at 4,000 kg/cm² for the other rocks. The dependence of velocity upon composition is briefly discussed.

INTRODUCTION

A dynamical method for the measurement of the rigidity of rocks at high pressure was described in an earlier paper;[2] results were published for a number of igneous rocks at pressures up to 4,000 kg/cm², at 30° C. and at 100° C. In the continuation of this work, the pressure range has been extended to 9,000 kg/cm², and the temperature range has been nearly doubled by its extension to 150° C. New measurements are given below for a few igneous rocks at the highest pressures and temperatures. Determinations to 4,000 kg/cm² are given for a number of sedimentary and metamorphic rocks, which have hitherto been very little studied at high pressure. Finally, rough correlations between velocity at high pressure and composition are shown graphically for certain important groups of rocks.

For all questions of experimental technique, the reader is referred to the paper cited above. The method consists in the determination of the resonant frequency of torsional vibration of right circular cylinders of the materials under investigation. The vibrations are excited and detected by electromagnetic devices not unlike telephone receivers. From this frequency F, the velocity V_S of waves of shear is found from the relation, $V_S = 2FL$, where L is the length of the

[1] Paper No. 69, published under the auspices of the Committee on Research in Experimental Geology and Geophysics and the Division of Geological Sciences at Harvard University.

[2] Francis Birch and Dennison Bancroft, "The Effect of Pressure on the Rigidity of Rocks," *Jour. Geol.*, Vol. XLVI (1938), pp. 59–87, 113–41.

cylinder. This velocity is related to the rigidity G and the density ρ by the expression, $G = V_S^2\rho$. These experiments may thus be described as the measurement of rigidity or of the velocity V_S. This velocity is the velocity of propagation of disturbances involving distortional strains alone (no compression); it is thus the velocity denoted in seismology as V_S or the velocity of the S waves.

A number of the specimens of sedimentary and metamorphic rocks were made from drill cores taken from bore holes along the line of a tunnel from the Wachusett Reservoir; for the use of these materials the writers are indebted to Mr. I. G. Crosby of Cambridge, and to the Metropolitan District Commission of Boston.

TABLE 1

MATERIALS STUDIED

Designation	Place of Origin	Density
Tonalite	Val Verde, Calif.	2.76
Albitite (Oligoclasite)	Sylmar, Pa.	2.62
Anorthosite (Bytownite)	Stillwater Complex, Mont.	2.74
Hypersthenite (Bronzitite)	Stillwater Complex, Mont.	3.27
Dolomite	Bethlehem, Pa.	2.82, 2.83
Limestone	Nazareth, Pa.	2.69
Hell Gate gneiss	Hell Gate, N.Y.	2.65
Massive pyrite	Noranda Mines, Quebec	4.85
Drill cores from the Boston region:		
Granodiorite	Weston (17.3)	2.63
Quartz-muscovite schist 1	Framingham (9A2)	2.70
Quartz-muscovite schist 2	Framingham (10.2)	2.73
Chlorite schist	Framingham (12)	2.95
Biotite gneiss	Solomon's Pond (A2)	2.91
Sandstone	Brookline (24.3)	2.61
Argillite	Brookline (23.13)	2.76
Slate	Everett (52.12)	2.67

DESCRIPTION OF MATERIALS

In Table 1 are listed the densities of the rocks and their places of origin, including the number of the bore hole and its approximate location for the samples obtained from the Wachusett Tunnel line.

Analyses of the tonalite,[3] albitite,[3] anorthosite,[3] hypersthenite,[2]

[3] Francis Birch and Harry Clark, "The Thermal Conductivity of Rocks and Its Dependence on Temperature and Composition," *Amer. Jour. Sci.*, Vol. CCXXXVIII (1940), pp. 529–58, 613–35.

dolomite,[5] and limestone[4] have been published; new analyses of thin sections are given below for the granodiorite, two specimens of quartz-muscovite schist, chlorite schist, biotite gneiss, and shale (argillite). These studies were made by Mr. R. L. Griggs of Harvard University. No analyses are available for the Hell Gate gneiss, slate, sandstone, or massive pyrite.

An error in the published analysis[3] of the anorthosite from the Stillwater Igneous Complex has kindly been brought to our attention by Professor H. H. Hess. The plagioclase is of the composition $Ab_{15}An_{85}$ (bytownite), instead of $Ab_{35}An_{65}$, as given in the former paper. Professor Hess also points out that the Stillwater "hypersthenite," of which he has published a chemical analysis,[6] would be more correctly described as "bronzitite."

Granodiorite.—A coarse interlocking aggregate of quartz and microperthite, holding subhedral tablets and blocks of the plagioclase.

MODE	Per Cent by Volume
Quartz	33.5
Microperthite ($Or_{50}Ab_{50}$)	30.0
Oligoclase ($Ab_{85}An_{15}$)	28.0
Magnetite	4.0
Muscovite and sericite	2.5
Carbonate	2.0
Apatite, zircon, sphene, leucoxene	0.5
	100.5

The grain size of the quartz and microperthite ranges from 3 to 0.5 mm. (average 2 mm.). The plagioclase grains tend to be somewhat smaller, ranging from 2 to 0.3 mm. (average 1 mm.). The microperthite shows a slight development of carbonate; the oligoclase is slightly clouded by the development of carbonate, sericite, and clay.

[4] W. A. Zisman, "Young's Modulus and Poisson's Ratio with Reference to Geophysical Applications," *Nat Acad. Sci.*, Vol. XIX (1933), pp. 653–65.

[5] Zisman, "Compressibility and Anisotropy of Rocks at and near the Earth's Surface," *ibid.*, pp. 666–79; J. M. Ide, "Comparison of Statically and Dynamically Determined Young's Modulus of Rocks," *Nat. Acad. Sci.*, Vol. XXII (1936), pp. 81–92.

[6] H. H. Hess and A. H. Phillips, "Optical Properties and Chemical Composition of Magnesian Orthopyroxenes," *Amer. Mineral.*, Vol. XXV (1940), pp. 271–85. Analysis 465 E 3h.

Biotite gneiss.—A medium-grained rock showing an intricate suturing and interlocking of the quartz and feldspar, the grains of which vary from 1 to 0.1 mm. Roughly parallel, ragged biotite leaves average 8 mm. long and 0.15 mm. thick.

MODE

Quartz	22.5
Oligoclase-andesine ($Ab_{70}An_{30}$)	53.0
Biotite	20.5
Sphene, clinozoisite, magnetite, apatite	4.5
	100.5

Quartz-muscovite schist (1).—A fine-grained, highly sutured mosaic of quartz and feldspar with parallel "stringers" of muscovite. In the quartz-feldspar mosaic the grains range from 0.8 to 0.1 mm. in diameter (average 0.25 mm.).

Quartz-muscovite schist (2).—The character of this rock is essentially the same as for the other mica schist; perhaps the quartz grains are not quite so well interlocked in this specimen.

MODE

	(1)	(2)
Quartz	57.0	59.0
Muscovite	19.0	20.0
Microperthite ($Or_{50}Ab_{50}$)	5.0	4.0
Oligoclase ($Ab_{71}An_{29}$)	10.0	....
Albite-oligoclase ($Ab_{89}An_{11}$)	....	9.0
Epidote	6.0	4.0
Chlorite, magnetite, carbonate	2.5	3.5
	99.5	99.5

Chlorite schist.—A fine-grained foliated rock made up of two distinct types of layers. One is an extremely fine aggregate of quartz and chlorite, in which the grains average 0.01 mm. in diameter, though some of the "leaves" of chlorite vary up to 0.2 mm. Within this layer there is a sprinkling of epidote. The other type of layer, much coarser, is composed of an interlocking aggregate of quartz and epidote grains varying between 1 and 2 mm. in diameter; associated with this layer are long sinuous aggregates of chlorite whose component plates average 3 mm. in length and 0.02 mm. in width.

The width of the two types of layer is highly variable, averaging about 3 mm.

MODE

Chlorite	36.0
Epidote	31.0
Quartz	27.0
Carbonate	4.0
Magnetite	1.5
	99.5

Shale (*argillite*).—A fine-grained, slightly metamorphosed shale composed of angular to subround grains of quartz, held in a matrix of extremely fine-grained sericite, chlorite(?), and clay(?) with the sericite predominating. The size of the quartz and feldspar grains averages 0.3 mm.; the matrix is too fine for a size determination.

MODE

Quartz	22.5
Sericite, chlorite(?), clay(?)	70.5
Oligoclase	2.0
Magnetite	4.5
	99.5

The new specimens of tonalite, albitite, anorthosite, and Hell Gate gneiss were roughly 8 inches long, $\frac{5}{8}$ inch in diameter; two cylinders were made of the gneiss, one with its axis perpendicular to the well-marked plane of foliation (designated "perp."), one with its axis in this plane (designated "par."). Three mutually perpendicular specimens were cored from a large block of the albitite. The hypersthenite specimen is one used in former work,[2] designated "hypersthenite 1." The dolomite and limestone specimens, also 8 inches long by $\frac{5}{8}$ inch in diameter, have been studied by Zisman at low pressure; the specimens of each of these rocks were mutually perpendicular in the original blocks.

The other specimens were all $\frac{7}{8}$ inch in diameter, ranging from 6 to 8 inches in length. The drill cores were originally 1.5 inches in diameter, and some were several feet long. Before these were ground to the size required for the pressure measurements, a few determina-

tions of velocity were made at 1 atmosphere on the original cores, untouched except for squared-off ends. The results are tabulated in the next section.

RESULTS

The new measurements are tabulated in Tables 2 to 5; in Figure 1 the velocity V_S is plotted as function of the pressure for a few of the specimens. Table 2 shows the velocity of shear waves at 30° C.

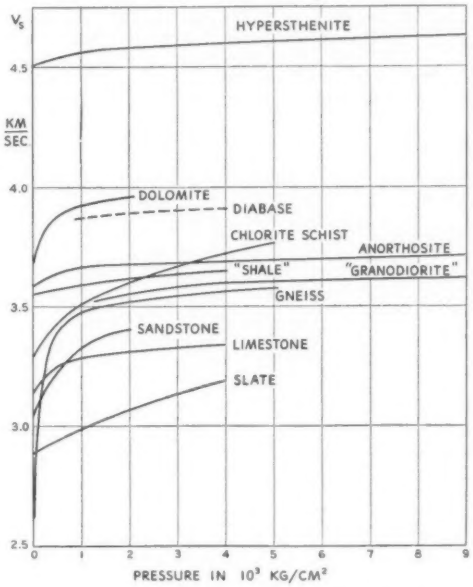

FIG. 1.—Velocity of shear waves in various rocks at high pressures

for the igneous rocks, except the albitite, as well as the pressure coefficients of velocity and of rigidity at pressures above about 6,000 kg/cm². The velocities in this table have been corrected for the change of length of the specimens under pressure; this correction has not been made for the other rocks, but it is probably not greater than 0.5 per cent for 5,000 kg/cm². The rigidity G may be found in terms of these velocities and the density ρ from the relation, $G = V_S^2\rho$. The coefficients giving the change of rigidity and of

velocity with pressure have also been corrected for the change of length and of density with pressure. Four figures are given for the sake of smooth differences, and the relative precision justifies this number of figures, but the absolute error on the velocity may be as great as 0.5 per cent.

The velocities in the other rocks are given in Table 3. Although the precision of any individual determination is as high for these materials as for the igneous rocks, the velocity is not always a single-valued function of the pressure; the curve of velocity versus pressure found on decreasing the pressure from its highest value will not in general coincide with the curve found on increasing the pressure. The difference between the two values of velocity at a given pressure is usually less than 1 per cent except at low pressure. The largest differences were found for the chlorite schist and the quartz-muscovite schist (1); for these materials, values are tabulated for increasing and for decreasing pressure. On successive runs, the amount of irreversibility tends to decrease. This effect has been discussed with reference to the character of the stress in such materials when they are exposed to pressure on the external surface alone, as in

TABLE 2

VELOCITY OF SHEAR WAVES AT 30° C.
(km/sec)

Pressure (kg/cm²)	Granodiorite	Tonalite	Anorthosite	Hypersthenite
1	3.14	3.12	3.59	4.50
500	3.45	3.45	.....	.....
1,000	3.507	3.554	3.669	4.566
2,000	3.556	3.599	3.675	4.576
3,000	3.575	3.627	3.680	4.585
4,000	3.580	3.643	3.684	4.593
5,000	3.583	3.650	3.688	4.598
6,000	3.586	3.654	3.691	4.602
7,000	3.589	3.658	3.694	4.606
8,000	3.592	3.663	3.697	4.611
9,000	3.595	.....	3.700	4.615
$10^6 \frac{1}{G}\frac{\Delta G}{\Delta P}$	3.5	4.4	3.0	2.9
$10^6 \frac{1}{V_s}\frac{\Delta V_s}{\Delta P}$	0.7	1.3	0.8	1.0

these experiments, where penetration by the pressure medium (nitrogen) is prevented by an impermeable sheath of copper foil. It is clear that the stresses on the individual crystals cannot be hydrostatic; with sufficient pressure, flow and fracture occur at the contacts between crystals.

Irreversible effects are also observed when the temperature is changed at constant pressure. Since the total change of velocity

TABLE 3

VELOCITY OF SHEAR WAVES AT 30° C.
(km/sec.)

Rock	Pressure, kg/cm²						
	1	500	1,000	2,000	3,000	4,000	5,000
Albitite 1	3.26	3.338	3.362	3.392	3.413	3.429	.....
2	3.30	3.337	3.362	3.387	3.402	3.412	.....
3	3.40	3.441	3.464	3.491	3.509	3.522	.....
Hell Gate gneiss:							
"Par."	2.49	3.29	3.47	3.55	.....	.....	.....
"Perp."	2.62	3.41	3.48	3.52	3.55	3.57	3.58
Biotite gneiss	3.42	3.51	3.54	3.57	.....	.....	.....
Chlorite schist:							
(Increasing P.)	3.30	3.45	3.52	3.60	3.66	3.72	3.76
(Decreasing P.)	3.37	3.52	3.57	3.65	3.71	3.74	.....
Quartz-muscovite schist 1:							
(Increasing P.)	2.92	3.42	3.57	3.69	3.74	3.76	.....
(Decreasing P.)	3.12	3.53	3.65	3.73	3.77	3.79	.....
Quartz-muscovite schist 2	2.76	3.42	3.57	3.63	3.66	3.68	3.69
Sandstone	3.04	3.24	3.34	3.40	.....	.....	.....
Slate	2.89	.....	2.98	3.07	3.14	3.19	.....
Argillite	3.55	3.57	3.59	3.61	3.63	3.65	.....
Limestone 1	3.15	3.27	3.29	3.32	3.33	3.34	.....
Limestone 3	3.14	3.27	3.29	3.32	3.33	3.34	.....
Dolomite 1	3.37	3.84	3.88	3.92	.....	.....	.....
Dolomite 2	3.67	3.89	3.92	3.95	.....	.....	.....
Massive pyrite	3.81	3.87	3.92	3.96	3.98	4.00	.....

for a change of temperature of 120° C. is not over 2 per cent, even a small irreversible change has a considerable influence upon the precision with which the temperature coefficient can be found. The temperature coefficients shown in Table 4 are average values; their uncertainties can be gauged in terms of the permanent increases of velocity after heating and cooling, which are also tabulated. When a rock is heated to 150° C. under pressure of 4,000 kg/cm² or more, its velocity decreases, but tends to drift toward higher values, in

marked contrast with the behavior on heating at 1 atmosphere.[7] It is not possible to say how great a change might be produced by an indefinite period of heating under pressure, nor what effect will be found on heating to still higher temperatures. It seems likely that the absolute change of velocity will be small, but the effect upon the temperature coefficient of velocity may be considerable. The error of the coefficients in Table 4 probably does not exceed 20 per cent so long as the temperature does not exceed 150° C. Corrections have

TABLE 4

EFFECT OF TEMPERATURE ON THE VELOCITY OF SHEAR WAVES AT HIGH PRESSURE (30° C. TO 150° C.)

Rock	Pressure (kg/cm^2)	$-10^6 \frac{1}{V_S} \frac{\Delta V_S}{\Delta T}$	Permanent Increase of Velocity (Per Cent)
Quincy granite	4,300	45	.....
Granodiorite	4,300	30	0.1
Tonalite	4,300 and 8,600	52	0.14
Albitite 2	4,300	77	.....
Albitite 3	4,300	85	.....
Anorthosite	4,300	45	0.1
	8,600	35	0.16
Hypersthenite	4,300	84	0.05
	8,600	62	(loop at 90°, 0.5)
Hell Gate gneiss, "par."	2,000	82	(no cooling curve)
Chlorite schist	4,300	136	0.23
Quartz-muscovite schist 1	4,300	79	0.4
Quartz-muscovite schist 2	4,300	65	0.2
Slate	4,300	156	1.8
(Iron)	1	150	.....

been applied for the thermal expansion of the specimens, using estimated values of thermal expansion; fortunately, this correction is relatively small. Measurements were made at three temperatures: 30°, 90°, and 150°; over this small range, the velocity was roughly a linear function of the temperature, except where irreversible effects appeared.

A striking example of exceptionally large "permanent" changes produced by compression is not given in the tables. A specimen was made from a stony meteorite, a hypersthene chondrite, containing

[7] J. M. Ide, "The Velocity of Sound in Rocks and Glasses as a Function of Temperature," *Jour. Geol.*, Vol. XLV (1937), pp. 689–716.

about 6 per cent of metallic nickel-iron. The original V_S was about 1.04 km/sec; on compression to 9,000 kg/cm², this velocity increased to over 3.6 km/sec, with no indication that the region of approximate constancy had been reached. This effect is an indication of unusual initial porosity; few terrestrial rocks show changes of velocity of more than 50 per cent on compression to a few thousands of atmospheres. A possible explanation would be that this meteorite had been heated to several hundred degrees at low pressure.

Measurements on a few of the rocks at 1 atmosphere and room temperature are given in Table 5. These include values of the velocity of plane compressional waves in rods (V_L), the velocity of shear waves, the velocity V_P of compressional waves in an unbounded

TABLE 5

ELASTICITY AT 1 ATMOSPHERE

Rock	Length (cm.)	V_P (km/sec)	V_L (km/sec)	V_S (km/sec)	σ
Granodiorite	31	5.00	4.78	3.10	0.187
Chlorite schist	35	4.97	4.89	3.27	.117
Quartz-muscovite schist 1	20	4.67	(4.49)	(2.92)	.181
Quartz-muscovite schist 2	20	4.24	(4.16)	(2.76)	.135
Slate	39	4.34	4.27	2.86	.115
Argillite	20	5.34	(5.28)	(3.55)	0.105

mass of the material, and Poisson's ratio σ. The two directly measured quantities are $V_L = (E/\rho)^{1/2}$, and V_S. The other quantities are calculated from these, on the assumption that the materials are isotropic. Measurements enclosed in parentheses were made on specimens which had been exposed to high pressure; the other values were determined upon larger specimens which had not been under pressure. The velocity V_P will also increase with the pressure; the relative increase will probably be somewhat greater than that of V_S.

The measurements on the oriented specimens of Hell Gate gneiss, dolomite, and limestone show that whatever may be the appearance of these materials, they are very nearly isotropic with respect to elasticity at high pressure. The differences of velocity in perpendicular directions, except at low pressure, are not over 1 per cent, even for the two specimens of the gneiss in which the foliation is very con-

spicuous. A similar effect has already been reported for gneiss from Pelham, Massachusetts. It is not possible to study the velocity in the drill cores except along the axis of the cores. A difference of velocity of about 3 per cent is found for the two specimens of quartz-muscovite schist, with almost identical compositions, but it is difficult to correlate this with schistosity. In these schists the foliation does not seem to preserve any preferred orientation; where it can be followed, it changes direction considerably in a single specimen. The albitite is not perfectly isotropic: the velocity in specimen 3 is uniformly about 3 per cent higher than in specimens 1 and 2. We have not yet been able to correlate this anisotropy with orientation of the albite crystals.

The hypersthenite (bronzitite) specimen (1) is the same one formerly studied to 4,000 kg/cm^2; the new measurements, made with a different oscillator and a new frequency calibration, are very close to those given before. The new value of the temperature coefficient at 4,000 kg/cm^2 is close to the former one (-76.10^{-6}) found for a different sample of the same rock; there is a somewhat smaller change with temperature at 8,600 kg/cm^2. For the anorthosite, also, the effect of temperature on the velocity is smaller at 8,600 kg/cm^2 than at 4,300 kg/cm^2. This behavior is qualitatively what we expect, but with higher temperatures and pressures, the numerical values of the coefficients may be substantially different.

Little work has been done with the sedimentary or metamorphic rocks at high pressure. It is evident from the present results that measurements at 1 atmosphere may be as misleading for these materials as for the igneous rocks. Large increases of velocity are found even for such apparently compact rocks as the schists and dolomite. The velocity in the dolomite is as high as in diabase or gabbro, and very much higher than in ordinary limestone. The replacement of calcium by magnesium clearly produces a structure of greater tightness and rigidity, as is evidenced also by the greater density of dolomite in spite of the lower atomic weight of magnesium, and by the higher thermal conductivity of dolomite. The velocities in the schists are somewhat higher than in granite. In these rocks occur large proportions of minerals whose elasticity is unknown, such as epidote and chlorite; for the micas, we have only the compressibility

of phlogopite mica.[8] The individual crystals of these minerals are probably highly anisotropic with respect to elasticity, but we can form an estimate of the mean values from the present work.

For this purpose, Figure 2 shows the velocity V_S at 4,000 kg/cm², for rocks composed principally of quartz plus feldspar, mica, chlorite, or epidote, plotted as function of the quartz content. These rocks include the quartzite, granites, and syenite (at zero quartz content) from our earlier paper, and the schists, granodiorite, tonalite, and argillite from the present paper. If the velocity in all of the minerals

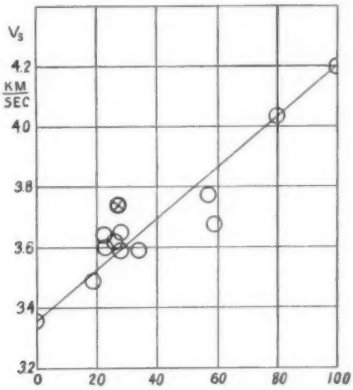

FIG. 2.—Velocity of shear waves at 4,000 kg/cm² in rocks composed of quartz, feldspar, mica, chlorite, and epidote, as function of quartz content (per cent by volume).

except the quartz were about equal, we might expect the points to fall on a smooth curve. Actually, except for the points for the chlorite-epidote schist and for one of the quartz-muscovite schists, the points lie very close to a straight line joining the point for the syenite and the point for 100 per cent quartz; the latter velocity was calculated by the use of Voigt's aggregate theory and the elastic constants for the quartz crystal. The syenite point should, perhaps, be placed a little lower, since it contains about 12 per cent of augite. The conclusion which we may draw from this evidence is that the micas, feldspars, clay and perhaps chlorite in these rocks differ little with

[8] L. H. Adams and E. D. Williamson, "The Compressibility of Minerals and Rocks at High Pressures," *Jour. Franklin Inst.*, Vol. XCCV (1923), pp. 475–529.

respect to mean velocity. The point for the chlorite-epidote schist (marked by a cross in Figure 2) is definitely high, suggesting a higher velocity probably for the epidote; the low value for one of the mica schists is not accounted for. All of the other points, including the one for the argillite, fall on the mean curve within the uncertainties of measurement of the composition. No such relation will hold for these rocks at low pressure. A different picture would result if the velocity of compressional waves were plotted against quartz content. Quartz is peculiar in possessing high rigidity with respect to its bulk modulus; to say this in another way, its Poisson's ratio is low. Whereas the velocity of shear waves in quartzite is higher than in syenite by about 20 per cent, the velocity of compressional waves (at high pressure) is nearly the same in the two materials. We might expect to find that V_P in all of the materials for which V_S is plotted in Figure 2 would be around 6.0 to 6.1 km/sec at 4,000 kg/cm^2.

A similar diagram can be constructed for rocks composed chiefly of plagioclase feldspar, pyroxene, and olivine. The velocity of shear waves in olivine rock is close to that in bronzitite, and we may assume for the moment that it will be nearly the same in the other varieties of pyroxene. In Figure 3, the velocity V_S at 4,000 kg/cm^2 is plotted against the content of pyroxene plus olivine for the norite, gabbros, and diabases of our earlier paper. The two points at zero pyroxene content represent the velocities in the anorthosite (bytownite) and the albitite (oligoclasite). The plagioclase in the basaltic rocks varies in composition between about $Ab_{50}An_{50}$ and $Ab_{30}An_{70}$. By a linear interpolation, using the velocities of the bytownite and the oligoclasite, we find that the velocities for the plagioclase in the other rocks should lie between about 3.56 and 3.64 km/sec. The smooth curve of Figure 3, drawn to 3.6 km/sec at the plagioclase side of the diagram, represents the measurements with nearly the accuracy of the determination of composition. Discussion of the shape of this curve will be left for another occasion. It is possible that the velocity in augite and diopside is lower than in bronzite.

Our studies of sedimentary rocks are still too few to form the basis of any sort of generalization, but it is already apparent that the velocities in these and in metamorphic rocks may overlap the

velocities in granites and even diabases. The problem of identifying materials in terms of seismic velocities alone may well be indeterminate, particularly at or near the surface—that is, at low pressures. The velocities appear to be determined primarily by two factors: composition and compactness. Once the porosity or "looseness" has been sufficiently reduced by compression, the composition alone is the determining factor. But a given velocity may be shown by rocks

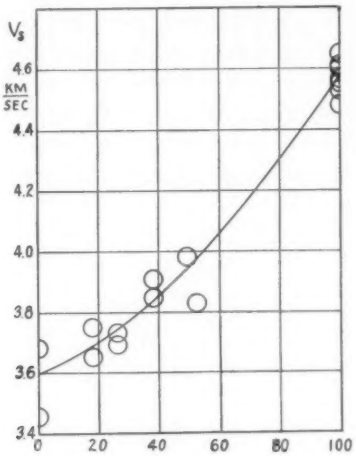

FIG. 3.—Velocity of shear waves at 4,000 kg/cm² in rocks composed of plagioclase feldspar, olivine, and pyroxene, as function of the content of olivine plus pyroxene (per cent by volume).

of very different compositions. Thus if V_S = 4.0 km/sec, we may be dealing with a quartzite with about 80 per cent of quartz, or with a gabbro with about 60 per cent of pyroxene and olivine. We could distinguish these if we had also the velocities V_P, which would be about 7.0 km/sec for the gabbro, as against about 6.1 km/sec for the quartzite. (All of these velocities are for 30° C. and 4,000 kg/cm²). But there would be little hope of distinguishing a compact shale containing about 25 per cent of quartz from a granite with the same amount of quartz.

On the other hand, the application of pressures of the order of a few thousands of atmospheres does not immediately eliminate the porosity of some rocks, and in some cases at least appreciably higher velocities might be found if compression could be continued sufficiently long. Heating, even to 150° C., speeds up the process of readjustment: a "permanent" increase of velocity of about 2 per cent was produced in the slate by heating to this temperature at 4,000 kg/cm^2. Higher pressures and temperatures may increase the velocities in such materials as limestones and sandstones to values near the limits set by the elasticity of the individual crystals. In nature, probably every degree of porosity from zero to the maximum consistent with coherence may be observed; the velocities will vary with the porosity, while the porosity itself will be determined not only by present depth but by the entire history of the formation.

THE PEAT DEPOSITS OF BERMUDA AND EVIDENCES OF POSTGLACIAL CHANGES IN SEA-LEVEL

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ABSTRACT

The evidences of recent submergence of Bermuda are from four sources: (1) topography, (2) submerged limestone caverns, (3) buried cedars and peat bogs, and (4) existing marshes.

The most convincing proof of postglacial submergence is found in the marshes on the island. The Hamilton marsh, 97.7 feet deep, gives evidence of a postglacial rise in sea-level of at least this amount. A test boring was made in this marsh to a depth of 42 feet. Five layers of woody peat were encountered, which are best interpreted as representing periods of relative stability of sea-level. They therefore show that the change was periodic and not gradual. They also offer an unusual means of determining the rate of peat accumulation and the length of postglacial time. It may be possible to correlate with their aid recent changes in level in other parts of the world.

INTRODUCTION

There is abundant evidence of Pleistocene and recent changes of sea-level on the island of Bermuda. Although considerable has been written on the subject, little is known about the details of these fluctuations. In 1937 the author made a few test borings in the Hamilton marsh and uncovered some new and interesting facts which seem to have an important bearing on this problem. The observations obtained from the borings are here recorded together with a discussion of their meaning and possible value in the study of sea-level changes in postglacial time. A brief review of the evidences which suggest that Bermuda was submerged as a result of eustatic changes in sea-level is included in the first part of the paper because these facts are closely related to the main problem.

QUESTION OF THE STABILITY OF BERMUDA

Tarr[1] and Verrill,[2] who have discussed the evidences of postglacial submergence of Bermuda, considered that the changes of position of

[1] R. S. Tarr, "Changes of Level in the Bermuda Islands," *Amer. Geol.*, Vol. XIX (May, 1897), pp. 293–303, Pls. 16–18.

[2] A. E. Verrill, "The Bermuda Islands," *Trans. Conn. Acad. Arts and Sci.*, Vol. XI, Part XI (April, 1902—Feb., 1903), pp. 413–956.

sea-level were caused by periodic sinking of the island as a result of forces operating within the earth rather than eustatic changes of sea-level caused by deglaciation of the continents. But they offered no evidence to support this assumption. Subsequent research has uncovered new facts which strongly suggest that the island of Bermuda has not moved either up or down to any great extent since the beginning of the Pleistocene glaciation and that the changes of sea-level on the island have been primarily eustatic.

Robert W. Sayles,[3] from his study of the eolian limestones and interbedded fossil soils of Bermuda, has shown that during the Pleistocene there were several periods of low- and high-water conditions on the island that can be correlated with eustatic or world-wide shifts in the level of the sea brought about by advances and retreats of the Pleistocene glaciers. During the periods of low sea-level, when Bermuda must have been much larger than now, circumstances were suitable for the accumulation of large areas of eolian limestone. On the other hand, during periods of high water like the present, conditions favored weathering and the production of fossil soils. From his study of the distribution of these eolian limestones or eolianites and fossil soils Sayles comes to the conclusion that there has been little or no vertical displacement of Bermuda since the beginning of the Pleistocene and that, therefore, submergence of the island was the result of eustatic changes in sea-level rather than subsidence of the land.

In 1912 a deep boring was made on the island in an unsuccessful attempt to get a source of fresh water for one of the hotels. The data obtained from this boring, which reached a depth of 1,278 feet below sea-level and revealed the volcanic nature of the core of the island, were recorded by Pirsson.[4] He states that eolian limestone was found from the surface to a depth of 245 feet below sea-level and that below this weathered volcanics were encountered to a depth of 455 feet. The eolian limestone is considered by Sayles to be of Pleistocene age and the volcanic material, which has been apparently re-worked by

[3] "Bermuda during the Ice Age," *Amer. Acad. Arts and Sci. Proc.*, Vol. LXVI, No. 11 (Nov., 1931), pp. 379–467.

[4] L. V. Pirsson, "Geology of Bermuda Island; the Igneous Platform," *Amer. Jour. Sci.*, Vol. XXXVIII (4th ser., 1914), pp. 189–206, 331–44.

the ocean, is much older.[5] The depth of the eolian limestone is strikingly close to the 75 meters, or 246 feet, estimated by Daly for the lowering of the sea during the maximum development of the Pleistocene glaciers.[6] It suggests that, if Daly's figure is at all accurate, there is further reason to believe that there has been no appreciable subsidence of Bermuda since the beginning of Pleistocene time.

TOPOGRAPHICAL EVIDENCE OF SUBMERGENCE

Evidences of the recent submergence of Bermuda are from four sources: (1) topography, (2) limestone caverns, (3) buried cedars and peat bogs, and (4) existing marshes.

According to R. S. Tarr,[7] submergence of the island of Bermuda in comparatively recent time is indicated by dune hills surrounded by salt water, the presence of drowned valleys, the irregular coast line, the abruptness of the coast on the south shore, and the disproportionate irregularity of the north and south coasts.

A. E. Verrill[8] has pointed out that parts of Harrington Sound are 80–95 feet deep and that many places in Great Sound, Murray Anchorage, and other localities have depths of 50–80 feet or more below sea-level. Believing that these depressions originated from collapsed subterranean channels and caves formed above sea-level, he thinks that they indicate submergence of at least 95 feet. He also noted that the outer reefs are covered with broken fragments of limestone and rough shelves of rock, which he considers to have been originally beaches and old shore ledges. If this interpretation is correct it would indicate a rise of sea-level of at least 150 feet. Verrill suggests that in a similar way the eroded Argus and Challenger banks now 180–240 feet below the surface, show submergence of that amount.

In this connection Sayles[9] points out that the Bermuda platform extending seaward from 1 to 3 miles is from 65 to 75 feet below sea-level and indicates a long period of erosion during which the

[5] *Op. cit.*, pp. 451–52, 389–90.

[6] R. A. Daly, *The Changing World of the Ice Age* (New Haven, 1934), p. 47.

[7] *Op. cit.*

[8] "Notes on the Geology of the Bermudas," *Amer. Jour. Sci.*, Vol. IX (4th ser.; May, 1900), pp. 324–28.

[9] *Op. cit.*, p. 447.

sea was that many feet below its present level. There is, however, no direct evidence from the topography itself to show when the changes in level occurred.

EVIDENCE FROM LIMESTONE CAVERNS

The limestone caverns located in several parts of the island, which have been investigated by A. C. Swinnerton,[10] show unquestionable evidence of submergence. Several of these caves are partly submerged, in some cases their floors being as much as 50 or 60 feet under water. Since he believes that limestone caves cannot be cut below sea-level, Swinnerton comes to the conclusion that the present position of the caves definitely indicates a rise of the sea of at least 60 feet. The presence of stalagmites many feet below the sea in these caves and the finding of stalactites at a depth of 30 feet in Hamilton Harbor bear out this contention.

The deposits found in some of these caverns suggest that the caves were being formed during the first period of low water, which Sayles has correlated with the Nebraskan stage of glaciation, and that sea-level was at that time at least 50–60 feet lower than now. A lack of appreciable solution of the cave deposits below water level indicates that the last submergence has been relatively recent, probably postglacial.

EVIDENCE FROM BURIED CEDAR TREES AND PEAT BOGS

In 1870, during excavations for a dockyard on Ireland Island, eolian limestone was encountered to a depth of 52 feet below sea-level.[11] At 46 feet a 2-foot layer of peat and red soil was uncovered in which were found vertical stumps of cedar trees. The exact age of this peat deposit is unknown. However, the Walsingham formation, which is considered to be in part of Kansan age, was said to have been found directly below the organic layer. This would suggest that the peat was probably middle Pleistocene in age. This peat deposit shows a definite change in sea-level of at least 46 feet since it was formed, but it does not give conclusive evidence as to the time when this change occurred.

[10] "The Caves of Bermuda," *Geol. Mag.*, Vol. LXVI (1929), pp. 79–84.

[11] J. M. Jones, "Recent Observations in the Bermudas," *Nature*, Vol. VI (Aug., 1872), p. 262.

Verrill[12] mentions the fact that submerged peat and cedar trees have been found in several of the harbors during dredging operations. In the channel of "the Reach" he himself dredged up peat from a depth of 15–20 feet. He considers these deposits to have been peat bogs and marshes like the Devonshire and Pembroke marshes, which were buried by the encroachment of the sea. He believes that these peat bogs probably came to be below sea-level through a long period of submergence before they were finally invaded by the ocean and this would explain why the peat was not worn away by wave action.

William Livingston,[13] former director of public works of Bermuda, also mentions a buried forest found during dredging operations in Town Cut Channel, St. Georges, 30–34 feet below sea-level. The trees were found in growing position in a peat layer 4 feet thick. In general the wood was in good condition, although some of the remains had the appearance of having been burned by some prehistoric forest fire.

Although a submergence of from 15 to 20 feet is indicated in the first instance and one of from 30 to 34 feet is suggested by the second, as in the other lines of evidence mentioned above there is no indication of the character of the change in sea-level or how recently it occurred.

EVIDENCE FROM EXISTING MARSHES

CHARACTER OF THE MARSHES

At the present time there are several marshes in Bermuda. The most extensive ones are the Devonshire marsh and the Pembroke marsh, situated in depressions between hills of eolian limestone in the central part of the island near Hamilton. The origin of these depressions is unknown, although possibly they are sink holes formed by the collapse of the roofs of underground caverns. The Devonshire marsh is over a mile long and half a mile wide, while the Pembroke marsh is three-quarters of a mile long and less than a quarter of a mile in width. Between them is a third nontidal marsh slightly smaller than the Pembroke marsh, located in a circular depression a short distance north of Hamilton.

[12] "The Bermuda Islands," *Trans. Conn. Acad. Arts and Sci.*, Vol. XII (1907), p. 82.

[13] "Evidences of Bermuda's Formation," *Mid-ocean*, Vol. XXV, No. 67 (Friday, Dec. 6, 1935), pp. 1–2.

These marshes, with the exception of the one near Hamilton, have been filled in or drained in recent years for the purpose of cultivation. Before the coming of man to Bermuda they were swamps which were covered with a thick vegetation including the Bermuda cedar (*Juniperus bermudiana* Linn.), the most abundant tree on the island, the palmetto (*Sabal palmetto* Michx.), bayberry or wax myrtle (*Myrica cerifera* Linn.), and a dense undergrowth of vines, shrubs, and several species of large ferns which reached a height of 6–8 feet.

The cedar apparently grew to a large size in these swamps, for Governor Roger Wood is reported to have sent to England in 1633 planks of cedar 30–32 inches wide and 12–13 feet long. It is said that logs of cedar as much as 5 feet in diameter are sometimes found buried in the peat. Because of the value of the wood the trees were cut off soon after the settlement of the island and the swamps were converted into marsh land.

The peat resembles that found in New England. It is of good quality and burns readily. It has been used on the island at times for fertilizer.

DEPTH OF THE PEAT

Borings were made by Governor Lefroy[14] in the Pembroke marsh in 1872, and he records its depth as 40–48 feet. The Devonshire marsh is reported to have a similar depth. Soundings made a few years ago in connection with the proposed filling of the Hamilton marsh gave a maximum depth of 97.7 feet.[15] This is, therefore, one of the deepest peat bogs on record in this part of the world. However, no evidence of the character of the bottom of the deposit was obtained.

In 1937 the author made a few test borings around the margin and near the center of the Hamilton marsh to determine the character of the peat. Using a Davis peat borer, samples were obtained at several horizons. Along the margins of the marsh the peat appeared to be only a few feet thick, but toward the center it was found to be considerably deeper. Here a depth of 42 feet was reached without striking bottom, but because of mechanical difficulties it was decided not

[14] J. Jones, *The Visitor's Guide to Bermuda* (Halifax, New York, and London, 1876), p. 121.

[15] C. H. Smith, director of Bermuda Department of Public Works, personal communication.

to go farther and so the maximum depth of the peat at this point was not determined. The borings showed that the sides of the basin in which the marsh is located are very steeply inclined, as would be the case if it were originally a sink hole. In this connection it is of interest to note that this depression has a depth similar to the depressions found in parts of Harrington Sound, which are considered by Verrill to be sink holes formed by the collapse of the roofs of underground caverns.

PHYSICAL CHARACTER OF THE PEAT

In the test boring near the center of the marsh the author encountered several horizons of woody peat. At the surface 2 feet of woody peat was found and other resistant layers were discovered at $10\frac{1}{2}$ feet, 19 feet, 25 feet, and 31 feet below the surface. The thickest layer, at least 2 or 3 feet thick, was found at the 25-foot horizon and considerable difficulty was experienced in penetrating it. Between these layers of woody peat and below the 31-foot level there was wet, pulpy peat easily penetrated by the boring apparatus. At no place was there evidence of an accumulation of wind-blown sand, and, since no such layers were reported in the deeper boring, it would thus indicate that at least the 42 feet tested and probably the entire 97.7 feet of peat were formed since Wisconsin time.

RESULTS OF THE MICROSCOPIC STUDY OF THE PEAT

The samples obtained at several horizons were examined under the microscope. The fragments of wood, pollen grains, and spores examined and identified showed that the entire thickness of the peat tested had accumulated in fresh or brackish water and that the elements of the vegetation had remained practically unchanged throughout the time represented by the deposition of the 42 feet of plant material. The woody layers were found to consist largely of cedar, rhizomes of ferns, and other woody plants.

Although pollen grains and spores were numerous, relatively few species of plants were represented. Fern spores were most abundant in all the specimens and included forms comparable to species of *Osmunda*, *Woodwardia*, *Pteris*, and *Aspidium*.

The pollen of *Myrica* was found in considerable abundance in all

the levels examined. The pollen of this tree is very similar to that of *Betula* (birch) and *Corylus* (hazelnut). However, the latter two genera are not found in Bermuda at the present time and probably never existed there. Therefore in this case there is no possibility of confusion in the identification as there may be where the three genera occur together. It has been frequently stated by students of pollen that the grains of *Myrica* are not preserved in peat deposits because they are nonresistant and easily destroyed. Although this statement has been challenged, no conclusive evidence has previously been presented to show that the pollen of *Myrica* can be preserved in peat. This is due to its similarity to *Betula* and *Corylus*, which frequently grow in the same region. But in the Bermuda deposits the unmistakable presence of *Myrica* pollen from the surface to the 42-foot level is definite evidence that these grains can be and are well preserved in peat for a considerable length of time.

The pollen of the palmetto was also found to be common in most of the samples. The greatest number recorded was from the 7-foot level where it constituted 31 per cent of the total number of pollen and spores. However, the percentages of this pollen varied considerably—but whether this has any significance cannot be determined from the few specimens that were studied. That the palms have existed in Bermuda for a long time and even during the glacial period is attested by the fact that fossil palms have been found in eolian limestone only a short distance from the Hamilton marsh. This would also indicate that the temperature of the island was not appreciably lowered during the Pleistocene.

Beside the foregoing, several species of grass pollen were found in all the peat samples, but the number per specimen was quite variable, averaging from 4 to 14 per cent of the total number of grains. Along with these, as in most peat samples, there was a relatively small number of pollen grains and spores, probably belonging to herbaceous plants growing in the bog association, that could not be identified.

Although the Bermuda cedar is the most common tree on the island and at one time was the principal element of the vegetation on the marshes, no pollen that could be referred to this tree was found

in any of the samples. This is not surprising, since it is a well-known fact that, because the pollen of this species soon breaks down in contact with water, it is therefore not preserved in peat deposits.

Because of the small number of specimens of peat obtained and examined no detailed pollen analysis was made, but until further data can be procured a few tentative conclusions can be drawn from this study. It was found that the woody layers mentioned above contained in general a large percentage of *Myrica* pollen, together with a great number of fern spores but very few grass pollen. This seems to indicate that during the time the woody layers were being formed the swamp was covered with cedar, with some palmetto, beneath which grew a dense growth of *Myrica*, ferns, and other plants. On the other hand, an examination of the interwoody layers revealed a considerable falling-off in the number of *Myrica* pollen and a corresponding increase in the pollen of grass. Apparently the swamp had become too wet for the growth of trees and shrubs and during these periods it was a marsh covered with grass and a few ferns, with cedar and palmetto growing along the margins. Thus, from the physical and botanical study of the peat, there is evidence of several changes in the character of the deposit, the causes of which are discussed below.

SIGNIFICANCE OF THE WOODY LAYERS

Because of the porous nature of the rock and the small size of the island, sea-level largely controls the water table, which in turn affects the marsh level. Consequently the top of Hamilton marsh (which is 1 foot above ordnance datum) is fresh, due to rainfall and runoff, but a few inches below the surface it becomes brackish. Any change in sea-level would modify the character of the surface of the marsh. Therefore this peat bog, which is not subject to tidal or wave influence, is very favorably situated for accurate measurements of sea-level changes. Thus it appears that this accumulation of peat indicates a definite rise of sea-level of at least 96.7 feet in recent, probably postglacial, time.

The layers of woody peat in the marsh may indicate either exceptionally dry climatic conditions, which seems unlikely, or periods when sea-level was nearly stationary over a considerable length of

time, during which the peat accumulation reached its climax and trees began to grow in abundance upon the surface. On the other hand, the thick layers of pulpy peat represent, according to this interpretation, times of rapid rise in sea-level resulting in a retrogression of the peat bog to wetter conditions, which brought about the destruction of the trees. Therefore, these five woody layers indicate that the postglacial change in sea-level was spasmodic and not a continual process.

However, the thickness of the woody layers would not necessarily indicate the length of unchanging sea-level, since it is doubtful that the accumulation of peat could always keep pace with a rapid rise in sea-level. The thickness would, therefore, depend upon the rapidity of the rise of the sea and the length of the following period of stability. Thus a rapid rise in sea-level followed by a short period during which the change was at a minimum would not give rise to a woody deposit. On the other hand, a slow period of submergence might produce a thick layer of woody peat. Such a woody layer could also be produced by a fall in sea-level resulting in the partial drying of the marsh, the thickness of the layer depending upon the amount of lowering. However, if sea-level on Bermuda was at any time 10–12 feet higher than at present, as some believe, there should be evidence in the marsh, since it would have been unquestionably flooded by the ocean. There is no evidence so far uncovered in the marshes themselves that suggests such a change.

Two of these woody layers may possibly be correlated with the deposits of submerged peat and cedar uncovered by dredging operations. Mention has already been made of peat and cedar being found in the channel of "the Reach" 15–20 feet below sea-level, and in Town Cut Channel 30–34 feet below sea-level. These levels closely correspond to the 19-foot and the 31-foot woody layers discovered by the author in Hamilton marsh. Whether these are of the same age is open to question. However, there is a striking correspondence in their relation to sea-level.

DEPTH OF THE PEAT AND POSTGLACIAL TIME

It is not known whether or not the 97.7 feet in the Hamilton marsh represent all or only a part of postglacial time. Daly believes

that the sea-level was about 75 meters or 246 feet lower than at present during the maximum phase of the Wisconsin glaciation. If so, this accumulation of peat seems on first thought to represent only a portion of post-Wisconsin time. But it must be borne in mind that the great increase in the size of the island due to a lower sea-level probably resulted in a rise of the water table in the center of the island. The slope of the water table during this time is not known, but it would depend upon several factors, including the amount of precipitation (which may have been greater than at present)[16] and the character of the substratum. There are no data on the character of the bottom of the peat deposit in Hamilton marsh. If the depression in which it is located is a sink hole, then it is probable that the peat rests upon impervious limestone, which would naturally result in a perched water table during the low-water stages. In any case the figured gradient of the water table would not be excessive.

From these considerations it appears that even if this thickness of peat represents the entire period of submergence or postglacial time, it indicates only the minimum amount of change in sea-level. However, in the later stages of the bog development, there would be a closer correspondence between marsh level and sea-level as the encroaching ocean reduced the size of the island.

EVIDENCE OF SUBMERGENCE IN MODERN TIME

The fact that 2 feet of woody peat occur at the surface and that cedars grew to large sizes in the marshes indicates that there has been no appreciable amount of submergence for several hundred years. W. N. Rice[17] and Verrill[18] compared the present coast line of Bermuda with maps of the island made in the seventeenth century. Verrill states that if any change in sea-level had taken place in recent time there would be a definite change in the size and shapes of the shallow bays and lagoons. No such changes were noted. Recent observations along the Atlantic coast[19] seem to substantiate these con-

[16] K. Bryan and R. C. Cady, "The Pleistocene Climate of Bermuda," *Amer. Jour. Sci.*, Vol. XXVII (5th ser.; April, 1934), p. 264.

[17] "Geology of Bermuda," *Bull. U.S. Nat. Mus.*, No. 25, Vol. I (1884), pp. 5–32.

[18] "Notes on the Geology of the Bermudas," *Amer. Jour. Sci.*, Vol. IX (4th ser.; May, 1900), p. 320.

[19] R. P. Goldthwait, "The Damariscotta Shell Heaps and Coastal Stability," *Amer. Jour. Sci.*, Vol. XXX, No. 175 (5th ser.; July, 1935), pp. 1–13.

clusions and indicate that sea-level has not changed perceptibly for several hundred if not several thousand years.

According to European workers the last period of stable sea-level before the present was about 5,000 years ago.[20] This was followed by a submergence ending about 1600 B.C., since which time sea-level has changed little on the British Isles at least. In North America along the Atlantic coast there is evidence to show that the last submergence followed a warm period which can be correlated with the Climatic Optimum which, according to leading authorities, occurred about 4000 B.C.[21] It therefore appears that 5,000 years for the last period of stable sea-level is quite logical, although the author from a study of New England deposits believes this is slightly too high and that 4,200 is a more accurate figure. It is quite likely that this last change in sea-level was world-wide and, therefore, would be reflected in the peat deposits of Bermuda. The $10\frac{1}{2}$-foot layer in the Hamilton marsh probably represents this former period of stability. If so, one may estimate the approximate rate of peat accumulation and the age of the marsh.

RATE OF THE ACCUMULATION OF THE PEAT AND AN ESTIMATE OF POSTGLACIAL TIME

If 2200 B.C. is assumed to be the approximate date of the period of stability represented by the $10\frac{1}{2}$-foot layer of woody peat in the Hamilton marsh, then the average rate of accumulation of woody and pulpy peat would be about 3 inches per century, which is a reasonable amount. At this rate it would take at a minimum about 39,000 years to accumulate the 97.7 feet of peat, which represents the maximum depth of the deposit. As was pointed out previously, there is no evidence to show that this thickness of peat represents all of postglacial time. In any case it must constitute most of this period, since the initial submergence of the island caused by the melting of the ice sheets was probably very rapid, and, therefore, sea-level would reach the -96.7 level in a relatively short time.

As in most methods of estimating geological time, the question of uniformity in rate is important. There are few data available as to the rate of accumulation of peat under varying circumstances. In

[20] C. Reid, *Submerged Forests* (Cambridge and New York, 1913), p. 115.

[21] C. E. P. Brooks, *The Evolution of Climate* (London, 1925), pp. 127–35.

most peat deposits, however, the rate of accumulation will depend largely upon climatic conditions and the type of vegetation involved. Fortunately, extremes in climate were absent during the formation of the peat deposits in Bermuda, which has an oceanic climate. Therefore the rate of the accumulation of peat was little modified by this factor and in this respect differs from continental deposits of the same nature. Furthermore, since the same species of plants are involved, the type of vegetation is not so important in its effect upon the rate of accumulation in the Bermuda deposits as it is elsewhere, where many types of plants are implicated.

There is another factor of importance that should be taken into consideration—namely, the amount of compression of the peat at a distance below the surface. In most peat bogs which the author has examined there is evidence of a considerable compaction of the peat with depth. In the Bermuda deposits, however, there is no sign of any great amount of compaction. Except for the resistant woody layers, caused by the presence of roots rather than compression, the peat was very loose, even in the lowest layers penetrated. This may be due to the buoyancy of the salt water and the presence of the mat-like woody layers made up of intertwining roots and stems, which, because of their resistant nature and because they have not changed their original position in the marsh during the passage of time, have thus acted as supports to the peat as it accumulated.

If all the foregoing factors are taken into account it appears that the marsh is somewhere between 40,000 and 50,000 years old, and if, as is likely, the accumulation represents all of postglacial time, then it gives an estimate of this time based largely upon sea-level changes.

Estimates of the length of postglacial time vary all the way from 7,000 years or less to 80,000 years and even more. Most estimates are admittedly mere guesses. The most reliable method of estimating this time so far presented is by the use of varves, but even it offers many uncertainties and is only of local application. Antevs,[22] using this method, believes that 40,000± years may be considered a good figure for the time since the Wisconsin glacier began to melt back in eastern North America.

[22] E. Antevs, "The Last Glaciation," *Amer. Geog. Soc. Research Series*, No. 17 (1928), p. 168.

Organic deposits have been utilized for determinations of time since the retreat of the continental ice, but in most places climatic factors have probably caused such a variation in rate that such estimates have been quite unreliable. However, in the case of the Bermuda peat deposits, variations of climate and other factors have been at a minimum, and therefore it is believed that these accumulations can be used to give a more reliable estimate than similar deposits in most other parts of the world.

CONCLUSION

The Bermuda marshes are unusual botanically and geologically. Because of special circumstances they indicate not only a considerable amount of submergence, but also apparently reflect several fluctuations or halts in the process during postglacial time. It may be possible with further study to date these periods and to correlate them with changes of sea-level and fluctuations of ice caps in other parts of the world. If this can be done, it would aid materially in unraveling the chronology of postglacial time. Several more woody layers may be present below the 42-foot level, and a complete examination of the deposit may reveal some other important facts.

It is believed that Bermuda is far enough removed from the areas of glaciation not to have been greatly affected by isostatic adjustment caused by waxing and waning ice caps. Therefore it offers a satisfactory and unusual picture of world-wide fluctuations of sea-level during Pleistocene and recent time.

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REVIEWS

Index of Paleozoic Coral Genera. By W. D. Lang, Stanley Smith, and H. D. Thomas. London: British Museum of Natural History, 1940. Pp. 231. 15*s*.

The present volume grew out of an attempt by Drs. Lang and Smith to revise critically Linnaeus' *Corallia Baltica* and to establish his species on a firm foundation. Inasmuch as it seemed essential to publish an index of Paleozoic genotypes as a preliminary to any complete account of the corals, Dr. H. D. Thomas was called in to collaborate in the work. It was not possible to complete the *Index* for press before Dr. Lang's retirement from the keepership of the geological department of the British Museum of Natural History. Thus the final revision has been largely in the hands of Drs. Smith and Thomas.

Dr. W. N. Edwards, who succeeded Dr. Lang as keeper of geology, in commenting on the work of the three collaborators, borrows from the editor of *Antiquity* the following remark: "The amount of work involved in tabulating facts is infinitely greater than in the composition of continuous narrative, the liability to error far greater, and the results, when published, misleadingly disproportionate in apparent volume to the labour of preparation." All students of paleontology who have had any experience whatever with the type of meticulous scientific compilation involved will agree that this is an appropriate garland to lay at the feet of the three authors.

The book is divided into five sections, as follows: (I) Introduction, (II) Acknowledgments, (III) "Genera," (IV) "Genotypes," and (V) "Literature." The Introduction, which was written by W. D. Lang, is little short of a paleontological masterpiece, and it should be "must" reading for all serious students of the subject. Some notion of the scope and character of this section of the work may be gained from the following quotations from it:

. . . . only by inch-by-inch methods of collecting, and by observing the evolution of separate characters of a population of allied forms through a detailed and consecutive series of beds, could any but a superficial view be gained of the evolution of a group of organisms. By this means it became possible to learn more about the origin of species than by any intensive study of Recent forms, because

while the whole present-day flora and fauna represent only one cross-section of the stream of life, and that an extreme section, paleontology deals with an unlimited number of successive cross-sections taken throughout that stream and each representing (in so far as the material has been found) a world-wide flora and fauna. Fundamentally, then, Recent zoölogy and botany are seen to be included in, and to form but a small part of, paleontology, and the full scope and nature of that science are at last appreciated.

The paleontologist ". . . . is embarrassed to find that the more abundant his material, and the more carefully it has been collected with due regard to its exact horizon, the more unconvincingly do the older classifications appear to express the phyletic relationships of the forms described."

. . . . Now, the systematist can only consider his lineages, or bundles of parallel or anastomosing lineages, as genera, and the terms of his linear series as species. But in adopting this outlook he is led into two difficulties; for where he applied names to the genera, he finds, as has already been pointed out, that the existing genera cut across his lineages, and are merely aggregates of species which belong to different lineages, but have reached the same grade of structure; and again, when he tries to diagnose what he considers to be the true genera, he will probably find that it is only possible to do so either on the basis of the genotype alone, or, if the diagnosis must be comprehensive, by expressing the relative fates of evolution of each character within the lineage—a diagnosis only practicable if expressed in a tabular form.

The genera are listed alphabetically (pp. 11–142). Included in the list are all genera and subgenera of Paleozoic corals, as well as other groups to which Paleozoic corals have been referred. Distinctions are made between those generic names which are properly formed according to the Rules of Zoölogical Nomenclature and those which are preoccupied, or are absolute synonyms, or are invalid for other reasons specifically stated; and the distinctions are shown by different kinds of type. Some idea of the enormity of the work involved in this compilation, so admirably completed by the collaborators, may be indicated partly by the fact that literally hundreds of genera are considered, and yet a single line of abbreviated comment under each name in itself may have involved weeks of detailed research.

The entire *Index* is up to date essentially through 1939, and the reviewer has noted no important omissions. R. Ludwig's 1865–66 *Corallen aus paläolithischen Formationen* received rough treatment from the authors, inasmuch as the fifty-six genera of Paleozoic corals erected in this

work are all suppressed in anticipation of a ruling by the International Commission on Zoölogical Nomenclature. According to the collaborators, the reinstatement of Ludwig's names would only cause confusion, and workers generally would refuse to recognize his terms. Although most students recognize the reasonableness of Lang and Smith's contention (1934), the statement that Ludwig's names are not to be used because of their length will not in itself stand the closest scrutiny, nor are his names rendered invalid by the "fanciful and unsound classification which he employs." Were these criteria alone sufficient for suppression, innumerable names in paleontology, as well as in zoölogy, would have to go by the board, and few authorities would be able to agree as to which ones should be retained.

The authors also attempt to settle the question as to whether Hall or Rominger has priority in regard to the issuance of their important works, both dated 1876. They cut the Gordian knot by proclaiming that Hall's work appeared first. This probability will meet with general approval, but the point actually is far from proved and possibly never can be proved to the satisfaction of all.

Genotypes (pp. 143–66) are arranged in alphabetical order of the trivial names. After each trivial name there follows, in order, the genus to which the species was originally referred, its author, the genus (in small capitals) of which it is the genotype, and the author of that genus. This is, of course, one of the most valuable features of the *Index*.

The literature is listed (pp. 167–231) in alphabetical order with the real date of publication given in square brackets after all dates which are not accurate. This bibliography will be invaluable to all students of the corals, for, so far as the reviewer has been able to discover, it is essentially complete. Not surprisingly, a few important American works, however, have been overlooked. Cases in point are Dr. Brandon Grove's significant papers on the Paleozoic corals in the *American Midland Naturalist*, some issues of which journal the collaborators obviously have seen. Wendell Sanford's excellent summary of the Tetracoralla in the *American Journal of Science* has also been missed.

Despite minor criticisms, which are possible, this *Index* represents the most important work on Paleozoic corals which has appeared since the issuance of Edwards and Haime's monumental publications of the middle of the past century. The paleontologists of a war-torn world will all be deeply grateful for the international scope of the treatment accorded the subject by the authors. Their thanks will also be due the British Museum

for issuing such an imperishable scientific study during a period when Britain is engaged in a death struggle. At such a time it is reassuring to read of somewhat earlier organic struggles, as reported by Lang:

> The genera with their attendant genotypes will enter, make their bow, and retiring to collect each its appropriate species, will reappear with them to unfold the evolutionary plot as formation succeeds formation until, in the Permian act, the curtain falls upon an unfinished drama. A coral fauna is there left in the midst of its evolution. Its future is an enigma, and we have no present knowledge as to whether complete extinction overtook it, or whether it underwent a rapid and radical reorganization, to reappear in Triassic times as a fauna of typical Hexacorals.

CAREY CRONEIS

The Devonian of Pennsylvania. By BRADFORD WILLARD. (Pennsylvania Geological Survey [4th ser.], Bull. G 19.) Harrisburg, Pa.: Bureau of Publications. 1939. Pp. 481; pls. 32; figs. 95.

In this excellent report on the Devonian of Pennsylvania the chapter on the Keyser limestone and Helderberg group was written by Frank M. Swartz, and that on the Oriskany group by Arthur B. Cleaves. The introductory chapter and those on the Middle and Upper Devonian correlations, geologic history, and paleontology were written by Willard. The Introduction gives an interesting historical résumé of the early work on Devonian, from Sedgwick and Murchison to the present Geological Survey of Pennsylvania.

The Devonian is the largest of the rock systems of Pennsylvania and, except for the Pennsylvanian, is greatest in exposed surface and in value of mineral products. This system has yielded much of the oil and gas of the state and most of the glass sand, as well as building-stones and a little iron ore. The formations are carefully described and illustrated with photographs and correlation charts, and details of selected sections are given in sufficient number to be of practical value to any geologist working in the area. There are thirty-one plates of fossils indicative of the various Devonian groups.

A few pages of epilogue are given which constitute a philosophical discussion of "the deplorable tendency to map time, not rocks." Willard prefers to limit the use of the term "formation" to mappable facies rather than contemporaneous beds.

LOUISE BARTON FREEMAN